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AUDUBON-ALBION STOCK, BOULDER COUNTY, COLORADO

BY ERNEST E. WAHLSTROM

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ABSTRACT

The Audubon-Albion stock in Boulder County, Colorado, is a composite intrusive consisting largely of monzonite and injected into and crosscutting pre-Cambrian granite, gneiss, and schist. The sequence is syenogabbro (oldest), monzonite, quartz-bearing monzonite, syenite, and granite. The monzonite probably formed from a syenogabbroic magma by crystal fractionation, but the younger differentiates, including quartz-bearing monzonite, syenite, and granite, resulted from more complex processes involving not only crystal fractionation but also assimilation or syntexis and transfer of materials by mobile, fluid-rich, alkalic solutions.

Age relations, the distribution of the various rock types, and the internal structures suggest that the stock was in part emplaced by repeated subsidence of a large conical block and in part by injection of magma into zones of weakness along internal or external contacts. There is little evidence to indicate that the walls of the stock were forced apart by the advancing magma.

Chemical analyses of the syenogabbro show an unusually high content of potash and are nearly identical with analyses of the basalt flows interbedded with the sediments of the Denver formation near Golden, Colorado. Correlation establishes the age of the oldest rock in the Audubon-Albion stock as early Eocene or, possibly, Paleocene. The exact age of the younger differentiates is uncertain.

INTRODUCTION

GENERAL STATEMENT

The Audubon-Albion stock is one of a series of petrographically related Tertiary igneous bodies intruded into and crosscutting the pre-Cambrian schists, gneisses, and granites of the Front Range of Colorado. Excellent exposures, sharp contacts, and well-developed internal structures permit accurate determination of the sequence and composition of individual intrusions within the stock.

Many attempts have been made in the past to outline the Laramide igneous history of the Front Range, but lack of detailed information in critical areas has not permitted accurate conclusions. The present paper attempts to supply data which will assist in the further development of a cogent and complete picture of the igneous activity that forms an integral part of the post-Cretaceous geologic history of the mountainous regions of north-central Colorado.

LOCATION

The Audubon-Albion stock is situated in the rugged mountains of the west-central part of Boulder County, Colorado, at about 40° 05′ latitude

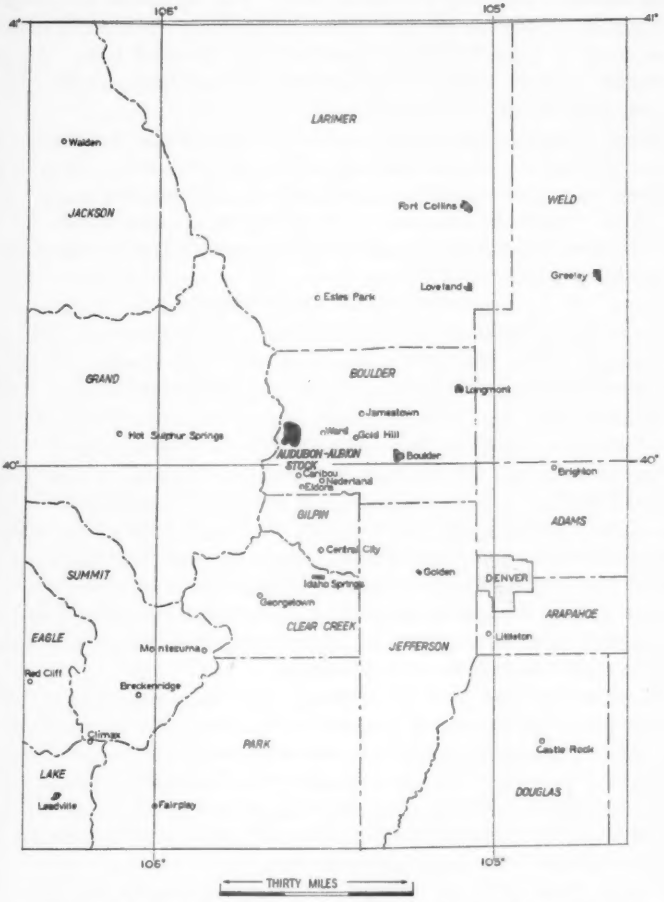

FIGURE 1.—*Outline map of north-central Colorado showing location of Audubon-Albion stock*

and 105° 35′ longitude (Fig. 1). The stock lies from 1 to 3 miles east of the northerly trending continental divide of the Front Range of the Colorado Rockies and is about 5 miles west of the small mining community

at Ward, Colorado. Long's Peak, in Rocky Mountain National Park, is approximately 13 miles due north of the center of the intrusion.

In plan, the stock forms a crude rectangle about 4 miles long in a north-south direction and nearly 2 miles wide. The northwest corner of the intrusion is exposed on the top of Mount Audubon whence the stock extends south to Lake Albion and east as far as Brainerd Lake. All these landmarks may be easily reached by horse or foot trails which originate at University Camp or Brainerd Lake.

The stock is near the northern end of the northeasterly trending mineral belt of Colorado. The coincidence of the mineral belt with a belt of Tertiary intrusions extending across the State is a fact long recognized by geologists. With the exception of the intrusion at Jamestown, the Audubon-Albion stock is the most northerly exposed Tertiary stock in the Front Range.

TOPOGRAPHY AND PHYSIOGRAPHY

The topography of the Audubon-Albion area is shown on the southeastern part of the Rocky Mountain National Park Quadrangle, mapped by the United States Geological Survey in 1912 to 1915 on a scale of 1/125,000. This map, although rather inaccurate, was enlarged and used as a base for the geologic map accompanying the present report.

The western half of the area is exceedingly rugged. Peaks ranging from 12,500 feet to more than 13,000 feet in altitude stand precipitously above deep glacial gorges and cirques. The total relief of the region is about 3000 feet; some of the canyons are as much as 2000 feet deep from rim to floor and have clifflike walls. In the eastern half of the area the glacial gorges become wider and shallower and are filled with thick hummocky deposits of glacial till. In contrast to the western half, lower rounded hills characterize the landscape.

Many of the high interstream areas near the continental divide are nearly flat or gently rolling. Accordance in level of these summit areas has led to the statement that they represent remnants of a former peneplain—the so-called "Flattop peneplain" of Lee (1922; 1923). Glacial moraines have locally concealed the eastern contact of the stock. Remnants of glacial material on the gently rolling top of Niwot Ridge north of Lake Albion cover a portion of the stock.

ACKNOWLEDGMENTS

The writer wishes to express his gratitude to the faculty of the Division of Geological Sciences at Harvard University for the assistance they unselfishly gave him in the preparation of the data and conclusions assembled in this paper. The writer is particularly indebted to Professor E. S. Larsen, who supervised the research in all its stages.

Acknowledgments are also due the staff of the Department of Geology at the University of Colorado, where full use of laboratory facilities was made for the chemical and microscopic analyses of specimens collected in the field.

REGIONAL STRUCTURE

GENERAL STATEMENT

The structure of the Audubon-Albion and adjacent regions is typical of the Front Range of Colorado. Two partly concordant pre-Cambrian batholiths have been injected into the extensive gneisses and schists of the Idaho Springs formation, and these, in turn, have been intruded by many crosscutting Tertiary dikes, small porphyry stocks, and larger stocks of coarse-grained equigranular rocks.

Faults are abundant but for the most part obscure because they are covered by talus or glacial till. Some faults are mineralized and are exposed in prospect holes and mines. In the area east of the Audubon-Albion region Lovering (1932) has mapped several northwesterly trending slightly mineralized fault zones; it is probable that some of these extend into the Audubon-Albion region. The faults are not described in the present work.

STRUCTURE OF THE PRE-CAMBRIAN ROCKS

Idaho Springs formation.—The oldest rocks in the region are the gneisses and schists of the Idaho Springs formation, which consists of foliated highly metamorphosed rocks largely derived from a thick series of early pre-Cambrian conglomerates, sandstones, shales, and impure limestones. The variety of rock types is great and includes predominant biotite-sillimanite gneiss and smaller amounts of garnetiferous gneiss, pyroxene gneiss, amphibolite, and quartzite; all gradations among the various rock types may be found. The foliation of the metamorphic rocks is emphasized in places by pegmatite and granite sills. The pegmatites are especially abundant near and at the contacts with younger granite masses.

The strike of the schistosity is almost uniformly a few degrees north of east, and the dip averages about 60 degrees to the north. Local exceptions to this rule are noted at the contacts of the gneiss with younger crosscutting rocks or where faults have displaced the rocks. The attitude of the planes of schistosity has exerted a controlling influence in the formation of many younger structures such as veins and flow layers in intruded granite. Masses of amphibolite derived from lime silicate layers in the Idaho Springs formation locally are abundant in the monzonite stock.

Boulder Creek gneiss.—The Boulder Creek gneiss (petrographically a granodiorite) forms a huge sill-like body which extends from Arapaho Valley into the eastern end of Albion Valley. The contacts of the sill

with the Idaho Springs formation are everywhere conformable, and, as the sill is traced along its strike, it gradually frays out into schist until it finally disappears.

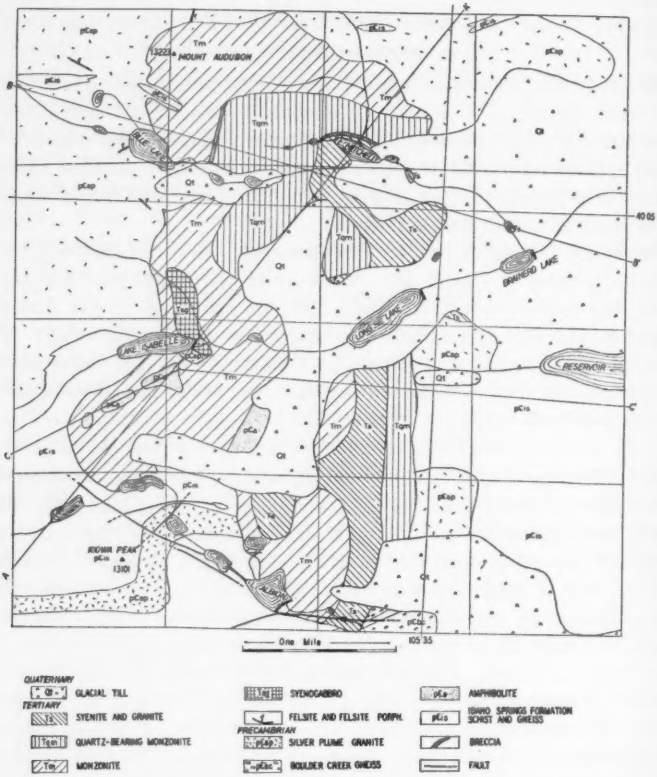

FIGURE 2.—*Geologic map of Audubon-Albion region, Boulder County, Colorado*

The gneiss has a pronounced structure paralleling the schistosity of the Idaho Springs formation. It is not certain whether its foliation is primary flow layering or whether it has resulted from reorientation of the constituent minerals by post-intrusive metamorphic stresses. The granite is locally porphyritic and in field exposures and thin sections shows granulation and slight rotation of the microcline phenocrysts. The attitude of the gneissoid structure agrees with what one might expect for an injected

body in the Audubon-Albion region, but the presence of granulated phenocrysts suggests that later stresses have contributed to the formation of the gneissic structure. The late stresses may have been manifestations of continuing intrusive energy after the almost complete solidification of the granite mass.

Silver Plume granite.—The Silver Plume granite—the most widespread igneous rock in the surrounding region—is exposed in large masses north and west of Mount Audubon and at Ward. Flow layering in this granite generally conforms with the schistosity of the Idaho Springs formation, but locally, where the granite crosscuts the gneiss, the flow layering roughly parallels the contact. In some places the flow structures in crosscutting bodies are haphazardly oriented and reflect the extreme irregularity of the contacts. The flow layering strikes, on the average, about N. 85° E. and dips about 60° N.

Inasmuch as the granite consists essentially of nearly equidimensional grains of quartz, tablets of orthoclase and microcline, with minor amounts of biotite, the flow lines are scarcely discernible. In a few places slightly elongate feldspar tablets trace out nearly horizontal flow lines.

Primary structures in the Silver Plume granite have exerted considerable influence in the development of faults and fractures. In field exposures the granite is highly jointed and breaks into large, flat, rectangular blocks. The best developed joints parallel the flow layering, and the less well-developed surfaces parallel respectively the linear structure and a plane at right angles thereto.

STRUCTURE OF THE TERTIARY ROCKS

Audubon-Albion stock.—The Audubon-Albion intrusion (Figs. 2, 3) is a composite monzonite stock, crosscutting all the older rocks. Five intrusive rock types are present, and it appears probable that considerable time elapsed between the successive periods of injection, for each of the phases of the stock was chilled against an earlier phase. It is likely that each period of intrusion was followed by a period of time sufficient to allow each molten injected mass to solidify and cool before the next younger phase was intruded.

Parts of the individual intrusions form roughly concentric arcuate belts, which, in some respects, resemble those of ring-dike structures such as are found in the White Mountains of New Hampshire (Billings and Williams, 1935; Chapman, 1935) and in Scotland (Bailey *et al.*, 1924). The structures in the Audubon-Albion stock do not obtain the perfection of symmetry displayed in some ring-dike areas. The cauldron subsidence theory appears plausible in explaining many of the observed features.

Dikes.—A few small dikes of felsite and felsite porphyry cut the pre-Cambrian rocks in the vicinity of the stock. No dikes were found in contact with the stock.

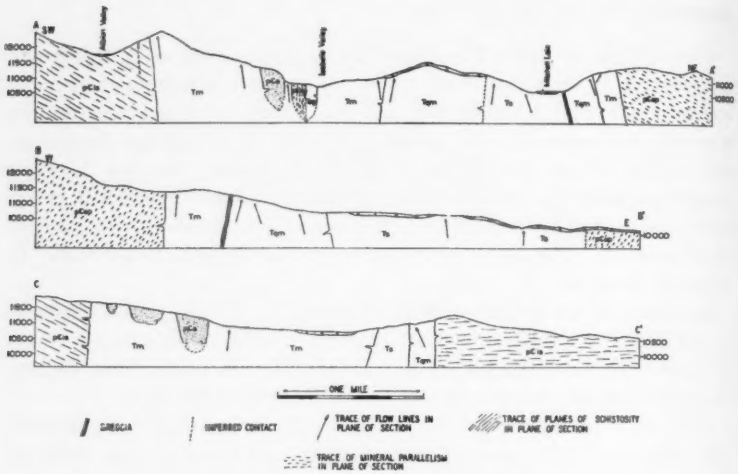

FIGURE 3.—*Geologic cross sections, Audubon-Albion region*

PETROLOGY OF THE STOCK AND RELATED ROCKS

PETROGRAPHY

Introductory statement.—The Audubon-Albion stock petrographically resembles stocks of monzonite exposed at Jamestown, Eldora, and Caribou, three mining camps in Boulder County, Colorado. At each locality rocks are found which resemble in every detail certain rocks in the Audubon-Albion stock. All the intrusions apparently are related genetically, and a study of one stock should yield information useful in the unravelling of the geologic history of the others.

Because the relief of the Audubon-Albion region is great, it is possible to observe the vertical changes in the rocks of the stock over a span of about 2500 feet. Excellent cliff exposures are found on the sides of the cirques and deeply glaciated valleys cut into the stock.

The contacts within the stock are sharp in most places and show intrusive relations and chilled zones. The sequence of intrusions, from oldest to youngest, is syenogabbro, monzonite, quartz-bearing monzonite, syenite, and alkalic granite. The Audubon-Albion intrusion contains, in addition

to the igneous rocks just named, veins rich in alkalic minerals and sulphides, which probably were formed during the last stages of igneous activity.

The oldest rocks of the stock were intruded into the area near its western border. Successively younger rocks were in general injected farther and farther eastward so that the stock now consists of several individual intrusions.

Syenogabbro.—The oldest intrusive rock in the stock is syenogabbro in a small mass just north and east of Lake Isabelle. The exposure does not exceed one-eighth of a square mile and is surrounded by monzonite. Inclusions of coarse amphibolite from nearby pre-Cambrian rocks are locally abundant. The syenogabbro is a finely granular massive, dark gray rock with an average grain diameter of about 0.75 millimeter.

Thin sections show that the texture is hypautomorphic-granular. In patches where large areas of pyroxene enclose random plagioclase plates the texture is ophitic. Some sections show phenocrysts of plagioclase. The following minerals, named in approximate order of crystallization, were identified: apatite, magnetite, plagioclase, pyroxene, orthoclase, and biotite.

Apatite and magnetite together constitute about 2 per cent of the rock by volume. The apatite is present in small, disseminated prisms and is included by later minerals. Magnetite euhedrons and subhedrons are poikilitically enclosed by plagioclase, pyroxene, and biotite flakes. Some fine-grained magnetite produces a schillerlike structure in the pyroxene.

The plagioclase is labradorite (An_{55}). Nearly half forms tabular phenocrysts approximately 2 millimeters in length; the remainder occurs in fine-grained aggregates, with grains about 0.5 millimeter in length. The phenocrysts are slightly zoned and have calcic cores grading into slightly more sodic feldspar toward the outside of the crystals; a few grains showing this gradation also display a calcic layer which lies outside the sodic layers.

The pyroxene is augite. It is replaced in patches by brown biotite and generally contains abundant magnetite grains.

Untwinned orthoclase appears in aggregates with plagioclase and forms anhedral grains which are molded around earlier minerals.

Biotite in abundant brown to light greenish-brown shreds and plates appears to have been the last mineral to form in the rock. The augite, in particular, contains small irregular patches, which suggest replacement after the augite crystallized.

The minerals are present in about the following percentages by volume: labradorite 55; augite 20; orthoclase 18; biotite 5; magnetite 1; and

apatite 1. The rock is classified as biotite syenogabbro. The presence of considerable orthoclase and biotite in a rock containing so much labradorite is noteworthy.

The norm of the syenogabbro, computed from a chemical analysis, does not agree closely with the mode. However, a slightly higher percentage of orthoclase in the norm is accounted for, in part, by the potash in the biotite of the mode. A rather large amount of titanium found in the analysis indicates that the abundant magnetite in the rock is titaniferous. An analysis and norm of syenogabbro from near the east end of Lake Isabelle are given in Table 1.

Monzonite.—Monzonite, the most abundant rock in the stock, is well exposed on Mount Audubon and at Lake Albion and forms, in plan, a strip from half a mile to a mile wide on the north, west, and south sides of the stock. Judging from the area of the exposures, the rock constitutes from half to three-fifths of the stock by volume.

Grain size, structure, and color vary from place to place. A contact facies near the top of Mount Audubon has an average grain size of about 2 millimeters; near the western border of the stock, in Albion Valley, large platy feldspar phenocrysts locally attain a width of 2 centimeters and a length of from 8 to 10 centimeters. Gneissoid to massive structures prevail.

In exposures the color ranges from dark gray to light gray. The shade of gray is a measure of the relative amount of ferromagnesian minerals and the extent to which the rock has been contaminated by assimilated or reworked inclusions of amphibolite and biotite gneiss. All gradations from angular unchanged fragments to ghostlike areas suggesting basic segregations are found. In general the monzonite is darkest gray where the inclusions are most abundant and lightest gray where they are rare.

The minerals identified in hand specimen are orthoclase, plagioclase, amphibole or pyroxene, and sphene. Both orthoclase and plagioclase are light gray and platy. The orthoclase shows Carlsbad twins and, where the grains are large, it contains abundant poikilitically included finely granular mafic minerals.

Black amphibole, either pargasite or hastingsite, in single grains or in clusters is distributed uniformly through the rock. Small lenses of coarse feldspar and radiating needles of amphibole 2 or 3 centimeters long are in places abundant, as on the south slopes of Mount Audubon. Many joint surfaces of the monzonite are coated with a thin layer of mafic minerals and sphene. The most abundant mafic mineral is black amphibole—either pargasite or hastingsite—which forms prismatic grains averaging about 3 millimeters in length. Joints in the granite and meta-

morphic rocks around the stock are locally mineralized by amphibole in the same manner as the joints in the stock. Probably the amphibole in both occurrences was introduced by hydrothermal solutions accompanying the stock.

Sphene in wedgelike yellow crystals 1 to 3 millimeters long constitutes about 1 per cent of the rock.

Microscopically the texture differs considerably in different specimens. Most of the variation is in grain size rather than in fabric. Hypautomorphic-granular and coarse porphyritic textures predominate. Plates of plagioclase commonly crystallized before anhedral orthoclase.

The following accessory minerals were identified: apatite, magnetite, sphene, garnet, pyroxene, amphibole, and biotite. Quartz is generally absent but is present in amounts up to 10 per cent by volume in a few thin sections. Secondary minerals are kaolin and "saussurite."

Apatite in prisms ranges from a fraction of a millimeter in the medium-grained monzonite to as much as a millimeter in length in coarse-grained varieties. Magnetite forms opaque anhedral to euhedral grains averaging about half a millimeter in diameter. The magnetite generally is most abundant in or near concentrations of other accessory minerals. Wedge-shaped euhedrons of sphene were observed in every thin section examined.

Light-green garnet is present in some specimens. A few irregular patches of garnet about a millimeter in diameter suggest partly formed blastocrysts developed by replacement of calcic plagioclase.

Pyroxene is an abundant accessory. Stubby, prismatic, poorly terminated crystals appear separately or in aggregates with amphibole, sphene, and magnetite. Much of the pyroxene has been irregularly uralitized to greenish-brown amphibole. In many slides the pyroxene is slightly zoned, and the grains display a faintly green core, which grades outward into a light-green rim. A poorly developed "hour glass" structure occurs in a few grains.

The optical properties of the pyroxene are as follows:

Indices	
α 1.702 ± .003	2V = 60°
β 1.711 ± .003	Optical orientation
γ 1.725 ± .003	X = b
	Center of crystal: Z∧c = 45°
	Rim of crystal: Z∧c = 48°

The index of refraction of the center and rim differs only slightly; the figures given are average values. From the optical properties the mineral appears to be diopside-hedenbergite (ratio of hedenbergite to diopside, 1:1; Winchell, 1933, p. 226), and a color gradation from very pale green to light green suggests an increase in iron content from the center of the

grains to the border, corresponding to a change from diopside-hedenbergite to a pyroxene slightly richer in hedenbergite or acmite molecules.

Pleochroic green amphibole is present in anhedral grains or in patches replacing pyroxene. Most grains average about a millimeter in length. The optical properties of the amphibole differ from specimen to specimen, but all the amphibole is probably pargasite or hastingsite. A greater extinction angle and a more intense pleochroism of some varieties of the amphibole may be accounted for by a higher iron content.

Small flakes and shreds of brown pleochroic biotite locally appear to have replaced the amphibole.

Quartz, as well as biotite and garnet, probably resulted from contamination of the monzonite. In general, quartz is most abundant in the monzonite where the latter is flanked by granite. Where the monzonite cuts across schist, quartz is generally absent or rare. The quartz is anhedral and makes a peculiar scalloped contact with the feldspar; many of the quartz grains have billowy convex surfaces. Locally, micrographic or myrmekitic intergrowths are abundant.

The plagioclase and orthoclase are present in about equal amounts. Most of the plagioclase forms euhedral or subhedral tablets with interstitial orthoclase or is poikilitically enclosed by orthoclase. The plagioclase is andesine ($Ab_{60}An_{40}$) grading into a border of andesine-oligoclase. A few grains of plagioclase have a thin outer albite layer. Small patches of albite-oligoclase are perthitically intergrown with the orthoclase (Pl. 1, fig. 1). The orthoclase grains are generally anhedral; however, in some facies of the monzonite the orthoclase forms subhedral tabular simply twinned crystals.

A type of alteration—"saussuritization"—does not depend on surface exposure but appears to have resulted from alteration by solutions contemporaneous with or slightly later than the period of injection and solidification.

The "saussurite" is present only in quartz-free monzonite. Partly because of this and partly because of the nature of the minerals composing the "saussurite," it seems probable that it formed by the alteration of nepheline. Nepheline was not found in any of the rocks examined.

The "saussurite" appears in interstitial aggregates or replaces andesine and orthoclase, in both of which it forms ragged, intersecting, or anastomosing veinlets. Some andesine grains are almost entirely replaced and consist of a few small isolated patches of unaltered feldspar embedded in "saussurite." Zeolite in the "saussurite" forms irregular feltlike aggregates of tiny radiating needles and is tentatively classified as thomsonite. With the zeolite are colorless shreds of sericite and, locally, pale-green

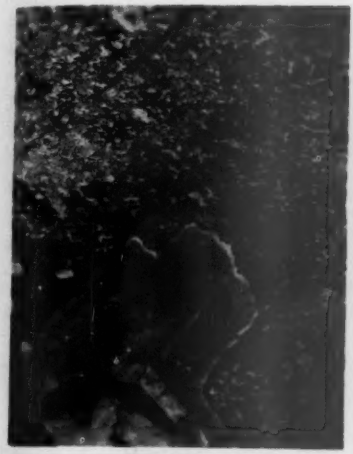

FIGURE 1. MONZONITE, NEAR LAKE ALBION

FIGURE 2. QUARTZ-BEARING MONZONITE, NEAR MITCHELL LAKE

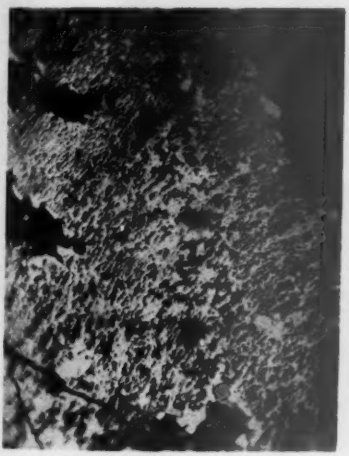

FIGURE 3. SYENITE, NEAR LONG LAKE

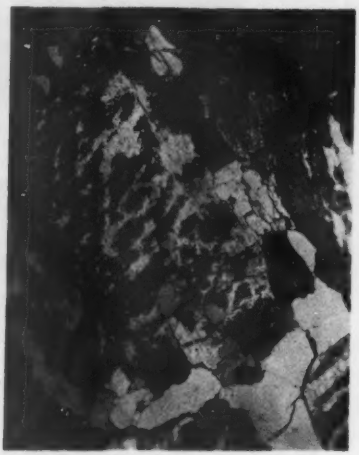

FIGURE 4. GRANITE, NEAR LONG LAKE

PHOTOMICROGRAPHS OF THIN SECTIONS FROM AUDUBON-ALBION STOCK

Showing progressive albitization of orthoclase. White and light gray = albite; gray and black = orthoclase; AN = andesine.

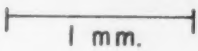

FIGURE 1. SCHLIEREN OF AMPHIBOLITE IN MONZONITE
Near Lake Isabelle. Note orthoclase introduced into amphibolite.

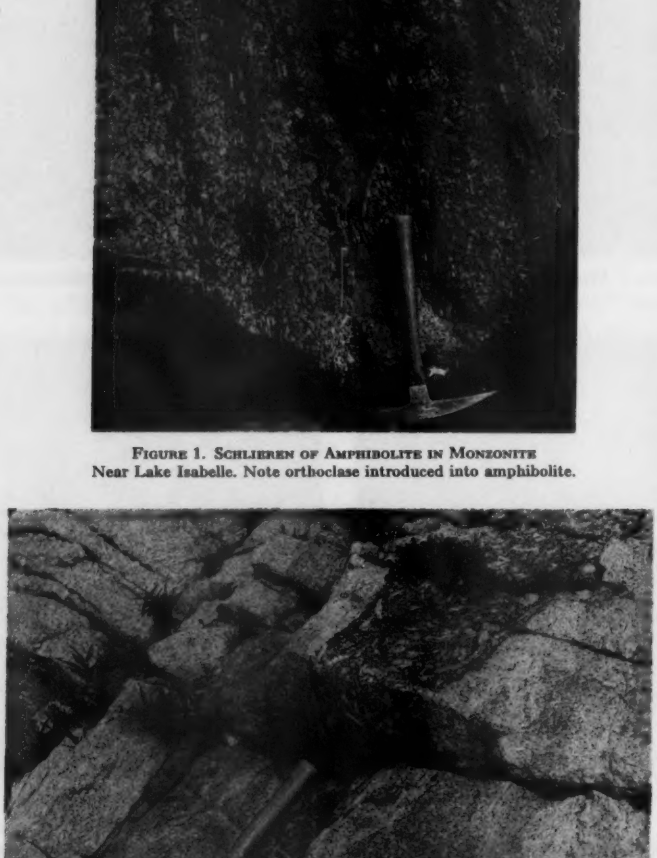

FIGURE 2. SCHLIEREN OF AMPHIBOLITE IN MONZONITE
Near Lake Isabelle.

SCHLIEREN OF AMPHIBOLITE IN MONZONITE

chlorite flakes. Colorless clinozoisite is present in small scattered grains in a few slides.

The monzonite has a considerable range in mineral composition. The following figures indicate the limits of variation from place to place within the stock; mafic minerals (pyroxene, amphibole, biotite, sphene, magnetite), 5 to 50 per cent; orthoclase, 25 to 75 per cent; quartz, 0 to 10 per cent.

Analyses of three samples selected from different facies of the monzonite are presented in Table 1.

Quartz-bearing monzonite.—The term "quartz-bearing monzonite" is applied to a light-gray, medium-grained monzonite which is well exposed in the valley of Mitchell and Blue lakes. It is also found on the ridge south of Lake Mitchell and is exposed in a strip about a quarter of a mile wide, extending south from Long Lake for a mile.

The area of exposure is about a third of that of the monzonite. The contact between the monzonite and the quartz-bearing monzonite is generally sharp, and apophyses and dikes of the quartz-bearing monzonite inject the monzonite; moreover, near the contact the quartz-bearing monzonite contains angular inclusions of the monzonite. On the south slopes of Mount Audubon and in the valley east of Blue Lake the contact is marked by a northerly trending breccia zone about 100 feet wide, which consists of angular fragments of monzonite in a matrix of porphyritic to fine-grained quartz-bearing monzonite. The matrix probably is a chilled contact phase of a magma which preceded the quartz-bearing monzonite now exposed but belongs to the same period of intrusion. The breccia zone is cut by dikes of quartz-bearing monzonite.

Fine-grained igneous material in the breccia suggests that a period of sufficient duration to allow for the cooling of the earlier monzonitic magma elapsed between the periods of intrusion of the monzonite and the quartz-bearing monzonite.

In exposures the quartz-bearing monzonite is light-gray, medium-grained, and contains few inclusions. It generally has three sets of well-developed joints which intersect approximately at right angles and cause the rock to break into large cuboidal or tabular blocks. The joints are controlled by platy and linear structure and are, in general, more closely spaced than those in the monzonite.

Orthoclase, plagioclase, amphibole, pyroxene, and sphene are recognizable in exposures. Carlsbad-twinned orthoclase and striated plagioclase have tabular habits and average from 3 to 5 millimeters in length. Crystals of yellow sphene are disseminated uniformly.

In thin sections the texture is typically monzonitic. The plagioclase is

present in two grain sizes: The larger grains are subhedral and accompany coarse interstitial orthoclase; the small grains are euhedral to subhedral and are poikilitically enclosed by orthoclase and microperthite. Micrographic and myrmekitic intergrowths of quartz and feldspar are common.

The following minerals were identified microscopically: plagioclase, microperthite (orthoclase and albite-oligoclase), and quartz as essential minerals, and magnetite, apatite, sphene, pyroxene, amphibole, and biotite as accessory minerals.

The tablets of plagioclase are zoned and have cores of andesine (An_{48}) which grade outward into rims of oligoclase (An_{20}).

The orthoclase has been partially replaced by albite-oligoclase, which forms irregular patches and gashes that generally appear to be haphazardly distributed through the orthoclase. The amount of plagioclase in this microperthite varies considerably from grain to grain and from one place to another in a single grain (Pl. 1, fig. 2).

Quartz forms small colorless interstitial grains and is present in all the thin sections examined.

Accessory minerals are not so abundant in the quartz-bearing monzonite as in the monzonite, but the optical properties are the same. Amphibole is relatively more abundant than pyroxene, and biotite is present only in sparse grains.

An estimate of the relative volumes by per cent of minerals computed from measurements of several sections follows: plagioclase, 40; orthoclase, 50; quartz, 4; mafic minerals, including pyroxene, amphibole, sphene, magnetite, and apatite, 6.

An analysis of a specimen (326) of quartz-bearing monzonite collected from an exposure half a mile southwest of Lake Mitchell is given in Table 1. A microscopic examination of the analyzed material showed that it contains only about 1 per cent free quartz, whereas the average quartz-bearing monzonite carries about 4 per cent quartz by volume.

Syenite and syenite contact breccia.—Light-gray to pinkish syenite is exposed southwest of Mitchell Lake, and a belt of this rock about a quarter of a mile wide extends south from Long Lake to the southern limits of the stock in Albion Valley. An excellent exposure of the syenite may be observed in a cliff in Albion valley an eighth of a mile north of Lake Albion.

The syenite constitutes about a fifth of the exposed part of the stock and is younger than either the monzonite of the quartz-bearing monzonite, for apophyses and dikes of syenite cut both the other rocks. Moreover, the contact facies of the syenite is finer-grained than the syenite in the center of the intrusion, suggesting that the syenite was chilled against the

quartz-bearing monzonite. In most places the contact is sharp, but at Lake Mitchell a wide brecciated contact zone has many peculiarities.

The syenite in many places is cut by numerous narrow dikes of pinkish quartz syenite and granite. Veins and stringers of quartz and alkalic pyroxene are locally abundant. The syenite is practically devoid of dark minerals, and in field exposures it has a dull, waxy appearance resulting from partial obliteration of the original texture by late albitization.

In exposures the rock consists almost entirely of medium-grained feldspar with a few prismatic grains of a black mafic mineral, probably amphibole. Near the contacts with the older rocks the syenite is porphyritic and consists of abundant feldspar phenocrysts in a sugary groundmass. Most of the syenite is nearly equigranular and has an average grain size of about 5 millimeters. In some thin sections the texture is hypautomorphic-equigranular; in others it is porphyritic-granular. Most of the grains are nearly equant.

The following minerals are seen under the microscope: plagioclase, microperthite (orthoclase and albite), quartz, amphibole, pyroxene, garnet, magnetite, and sphene. A few slides contain secondary zeolite and sericite.

The syenite contains a few scattered euhedral grains of oligoclase (An_{20}). Some of these are rimmed by a more highly birefringent albite layer. Most of the plagioclase is calcic albite and has irregularly replaced the orthoclase to form microperthite. A few large grains of orthoclase show differing degrees of replacement from one place to another in a single crystal. Moreover, in some specimens the microperthite is more than 50 per cent albite by volume, and, where this is true, the original texture of the rock has been largely destroyed. The resulting microperthite consists of ragged interlocking and sutured patches of orthoclase and albite (Pl. 1, fig. 3).

Quartz is rare in the syenite; however, a few tiny interstitial grains were observed in some specimens. Amphibole, pyroxene, magnetite, and sphene constitute about 1 per cent of the rock and are similar to the mafic minerals in the monzonite. One specimen contained a few grains of green isotropic garnet. Zeolite, probably thomsonite, and sericite occur in scattered patches in the syenite from Albion Valley. It is probable that the thomsonite and sericite are alteration products of nephelite, but nepheline was not identified in thin sections.

An analysis and norm of a slightly kaolinized specimen of syenite from Albion Valley are given in Table 1.

At Lake Mitchell the contact of the syenite with the older rocks is marked by a breccia zone which deserves special mention. The zone is about a quarter of a mile long and 100 to 200 feet wide, trends east, and borders the lake on its north and west sides.

The breccia consists of sheared and jointed felsite and felsite porphyry which contain abundant round to angular inclusions of igneous and metamorphic rocks. Quartz-bearing monzonite is especially abundant. Near the western edge of the zone the breccia contains large inclusions of dark-gray to black fine-grained metamorphic rock that at first glance suggests phyllite. The material appears massive in fresh specimens but on weathered surfaces displays foliation. Microscopically the rock is composed of alternate layers of fine-grained quartz and pleochroic green and bluish-green amphibole grains about 0.01 millimeter in length. The rock has been intensely fractured, and many of the laminae terminate abruptly against minute fault surfaces. Coarse-grained granite and pegmatite inclusions are locally abundant and probably came from the Silver Plume granite batholith.

With the exception of the quartz-bearing monzonite, none of the materials in the inclusions is found anywhere in the immediate vicinity of the breccia zone. The inclusions are probably either stoped from above or carried from depth by an ascending magma.

The main body of syenite is younger than the breccia. The contact between them is generally sharp, but in a few places it is gradational. The breccia is probably a chilled phase which slightly preceded the intrusion of the main body of syenite.

Microscopically the felsite and felsite porphyry in the breccia and the syenite are unlike. The felsite is much finer-grained and locally is rich in quartz and amphibole. Amphibole closely resembling the amphibole of the fine-grained schist and large corroded quartz grains suggest that both were introduced by assimilation of inclusions. This inference is supported by the fact that the distribution of quartz and amphibole in the felsite is very irregular.

Locally, joints and faults in the breccia have been mineralized by sulphides.

Granite.—Granite dikes and quartz veins are abundant in the syenite exposures on the ridge immediately north of Long Lake. The dikes range from a few inches to a few feet in width and generally are closely associated with narrow quartz stringers. In places the quartz stringers cut and make sharp contacts with the granite; elsewhere they grade into the granite and appear to have replaced it. Although the dikes and quartz stringers are numerous, their total volume compared to the syenite is insignificant. Accordingly, they were not mapped separately and are not distinguished from syenite on the geologic map.

In exposures the granite is light pink to gray and consists almost entirely of quartz and feldspar grains which average about 2 millimeters

in diameter. Pink feldspar forms small tablets or nearly equant grains. Quartz is present in interstitial glassy grains. The texture is hypautomorphic-granular.

Apatite, sphene, pyroxene, orthoclase, albite, and quartz were observed under the microscope. The apatite and sphene occur in small sparse euhedrons. The pyroxene is yellowish green, slightly pleochroic and forms small subhedral masses and is probably aegirite-augite. It makes up about one per cent of the rock by volume.

Euhedral orthoclase containing albite patches is conspicuous (Pl. 1, fig. 4). Some grains of orthoclase contain 50 per cent albite by volume, leading to the conclusion that the albite is of replacement origin and did not form by exsolution. The replacement probably took place during the last stages of the solidification of the magma. The patches and veinlets of albite are lenticular and, between crossed nicols, give the feldspar a striped appearance. Colorless, unstrained quartz is interstitial to the feldspar.

The minerals are present in the following amounts by volume: accessory minerals, including apatite, sphene, and pyroxene, about 2 per cent; feldspar (perthite), about 88 per cent; and quartz, about 10 per cent. The rock is granite, low in quartz, and approaching quartz syenite in composition.

The analysis and norm of a specimen of granite collected from a dike in a syenite exposure about a quarter of a mile north of Long Lake is included in Table 1.

Dike rocks.—A few small dikes, ranging in composition from rhyolite to dacite, are found in the Audubon-Albion region. Inasmuch as the dikes are not in contact with the stock, the relative age could not be determined. The dikes, chemically and mineralogically unlike the rocks of the stock, probably were injected considerably before or after the main period of intrusion.

Megascopically the dike rocks generally are light to medium-gray porphyritic felsites. The phenocrysts are quartz, orthoclase, plagioclase, and, in a few specimens, amphibole. The groundmass is very fine-grained but entirely holocrystalline.

Microscopically the following minerals were identified: magnetite, hornblende, biotite, plagioclase, orthoclase, and quartz.

Late alkalic differentiates.—The last event in the igneous history of the monzonite stock was the formation of veins and replacement deposits of sulphides. Associated gangue minerals are soda asbestos, aegirite, quartz, calcite, fluorite, barite, and potash feldspar.

TABLE 1.—*Rock analyses of the Audubon-Albion region, Boulder County, Colorado*

(E. E. Wahlstrom, Analyst)

	453	218	65	401	326	120	230
SiO_2	50.37	53.14	56.62	56.88	61.73	62.27	68.28
TiO_2	1.53	0.95	0.92	1.02	0.43	0.12	0.14
Al_2O_3	17.38	21.75	21.39	19.78	20.50	21.70	16.44
Fe_2O_3	1.91	1.35	1.18	2.94	0.61	0.61	0.77
FeO	6.06	2.79	2.18	2.39	1.45	0.41	0.42
MgO	6.16	1.58	0.29	0.63	0.60	0.04	0.02
CaO	8.25	7.14	5.64	5.88	3.20	1.94	0.58
Na_2O	3.16	4.46	4.91	4.69	6.21	5.84	3.70
K_2O	3.60	5.07	5.63	5.00	5.06	6.86	9.66
H_2O+	0.43	1.27	0.50	0.01	0.20	0.44	0.06
H_2O-	0.14	0.05	0.06	0.12	0.06	0.09	0.11
MnO	0.13	0.08	0.06	0.15	0.10	trace	0.06
P_2O_5	0.54	0.32	0.36	0.30	0.07	trace	trace
CO_2	0.08	0.15	0.27	0.30	trace	trace	0.05
Total	99.74	100.10	100.01	100.09	100.22	100.32	100.29

C. I. P. W. norms

or	21.13	29.47	33.36	29.47	29.47	40.59	56.71
ab	22.01	18.86	32.49	39.30	50.83	45.06	30.92
an	22.80	24.46	19.74	18.07	13.34	9.45	.....
ne	2.27	10.22	4.83	.....	0.85	2.27	.....
qu	.....	.....	.....	0.48	.....	.....	9.43
ap	1.34	0.67	0.67	0.67	.....	.....	.....
il	2.89	1.67	1.67	1.98	0.76	0.15	0.30
mg	2.78	1.22	1.86	4.18	0.93	0.93	0.70
di	11.77	7.48	4.21	4.45	2.07	.....	0.25
wo	.....	.....	.....	0.23	.....	.....	1.04
ol	12.07	4.22	.....	.....	1.52	.....	.....
ca	.....	.....	.....	0.68	.....	.....	.....
cor	.....	.....	.....	.....	.....	1.22	Ac 0.46
Sal	68.21	83.01	90.42	87.32	94.49	98.59	97.06
Fem	30.85	15.26	9.01	12.19	5.28	1.08	2.75

453. Syenogabbro. Shoshonose (II".5.3.3"). East end of Lake Isabelle.
218. Nepheline (?) monzonite. Shoshonose ((I) II.5(2)."3.3"). Trail, half a mile east of Lake Isabelle.
65. Monzonite. Pulaskose (I". 5.2" .3"). Eureka crosscut, Camp Albion.
401. Monzonite. Pulaskose (I(II).5.2".3"). Cirque in Mount Audubon.
326. Quartz-bearing monzonite. Larvikose (I.5.2.(3)4). Half a mile southwest of Lake Mitchell.
120. Syenite. Pulaskose (I.5.2".3"). Cliff north of Lake Albion.
230. Alkalic granite. Hettose (?) (I.(4)5.1.2(3)). Cliff north of Long Lake.

INTERNAL STRUCTURES OF THE STOCK

Inclusions (Pl. 3, fig. 1; Pl. 4, fig. 1) and schlieren (Pl. 2) are abundant, especially near the contacts and near screenlike masses. The inclusions locally attain a diameter of 100 feet. Three large masses of amphibolite

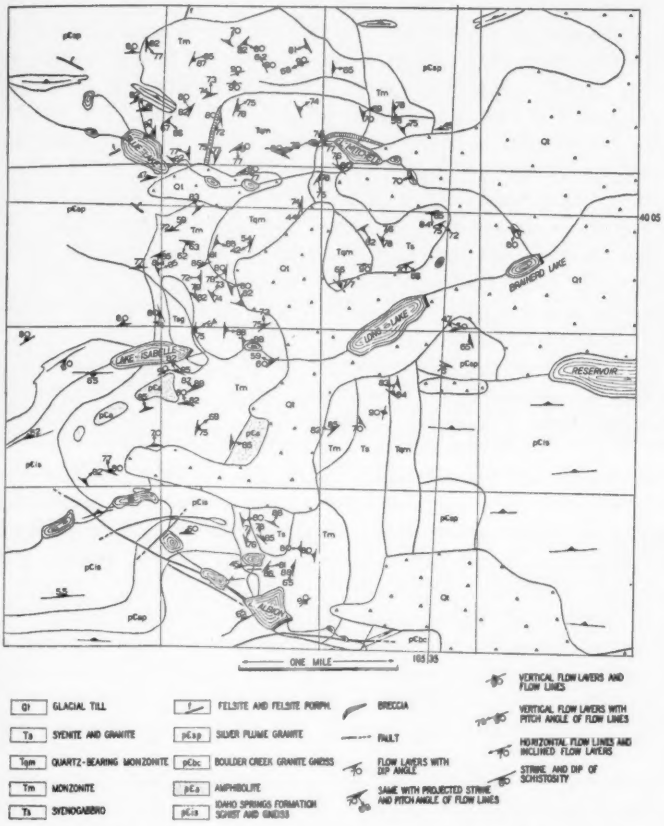

FIGURE 4.—*Map showing internal structures of Audubon-Albion stock and surrounding country rocks*

near Lake Isabelle (Pl. 4, fig. 2) are probably roof pendants. The inclusions exhibit all stages of assimilation. Some are angular; others are dragged out into schlieren. Tablets of orthoclase introduced into many schlieren are parallel to the feldspar plates in the surrounding monzonite.

Flow lines and flow layering generally parallel steeply dipping contacts (Fig. 4). The flow lines lie in the planes of the flow layering and ordinarily pitch steeply. Apparently the outermost walls of the stock are nearly vertical, but the younger intrusives within the stock dip steeply outward.

Three or more well-developed joint systems characterize the monzonite and quartz-bearing monzonite. The best developed system parallels the platy structure or flow layering; another consists of nearly vertical joints at right angles to the flow layers; and a third system, nearly flat, has formed almost at right angles to the flow lines. Joint surfaces locally are coated by streaks of hydrothermal amphibole, pyroxene, and pyrite dragged out by horizontal movement.

Some exposures exhibit two or more systems of diagonal joints. Apparently these are not related to any visible primary structures. Some of them have slightly displaced the joints paralleling the flow layering shortly after the intrusion of the quartz-bearing monzonite. Near Lake Albion the diagonal joints contain low-grade sulphide ores and are related to a brecciated fault zone cutting the stock.

Lenticular masses of brecciated rocks crop out near Lake Mitchell and just east of Lake Mitchell. These suggest the screens commonly associated with ring dikes.

DIFFERENTIATION AND IGNEOUS SEQUENCE

Correlation between mineralogic changes as observed under the microscope and chemical changes in differentiates as shown by variation diagrams aids in the determination of the mechanism of mechanisms of differentiation. Useful variation diagrams are those of Harker (1909), Von Wolff (Johannsen, 1931), and Larsen (1938).

A generalized Harker variation diagram of the Audubon-Albion stock is shown in Figure 5. One of its most useful features is its adaptability to computations of the compositions of hypothetical materials that might be added or subtracted during differentiation. Harker diagrams and computations used to study differentiation have been described by Bowen (1928).

The oldest exposed differentiate from the syenogabbro is monzonite, which, although present as a single continuous mass, displays local gradations in composition. The local presence of abundant schlieren and inclusions indicates that some of the variations have resulted from contamination. However, the greater part of the monzonite is comparatively free from inclusions and probably formed by crystal fractionation rather than by addition of foreign material to syenogabbroic magma. Computations on a Harker diagram indicate that the monzonite is derived from the syenogabbro (specimen 453) either by subtraction of material approxi-

FIGURE 1. WORKED-OVER INCLUSION OF AMPHIBOLITE IN MONZONITE
Near Lake Isabelle.

FIGURE 2. JOINT BLOCKS IN QUARTZ-BEARING MONZONITE
Half a mile southeast of top of Mount Audubon.

INTERNAL STRUCTURES OF MONZONITE

FIGURE 1. BRECCIA AT CONTACT BETWEEN AMPHIBOLITE AND MONZONITE
Near Lake Isabelle.

FIGURE 2. ROOF PENDANT OF AMPHIBOLITE IN MONZONITE
Southeast of Lake Isabelle.

AMPHIBOLITE MASSES INCLUDED IN MONZONITE

mating a mixture of anorthite, ilmenite, diopside, hypersthene, and olivine or by the addition of a mixture of orthoclase, nephelite, plagioclase, wollastonite, and magnetite. The latter material has an unusual composition and does not mineralogically resemble any of the wall rocks of the

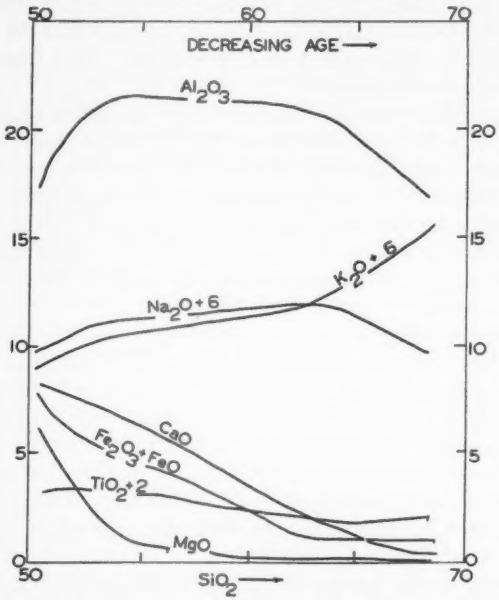

FIGURE 5.—*Generalized Harker variation diagram of rocks in Audubon-Albion stock*

stock nor does it suggest any known abundant rock. Hence, the monzonite probably resulted from the injection and solidification of a magma which formed as a result of a subtraction of early crystals of calcic plagioclase, pyroxene, olivine, and titaniferous iron oxides. Microscopically the syenogabbro contain pyroxene, plagioclase (An_{55}), orthoclase, biotite, and small amounts of magnetite, and apatite. Compared with the syenogabbro, the monzonite contains a much lower proportion of mafic minerals, plagioclase, sphene, and magnetite, an aggregate which chemically resembles the mixture of mafic minerals thought to have been subtracted from the syenogabbroic magma to produce monzonite. Accordingly, the conclusion reached by a study of the variation diagram is confirmed by the petrographic evidence.

After the solidification of the monzonitic magma, a considerable period elapsed, for it chilled the next younger intrusion—the quartz-bearing monzonite. The quartz-bearing monzonite is similar mineralogically to the monzonite except for a lower proportion of mafic minerals and grains of interstitial quartz. Thin sections of the quartz-bearing monzonite provide evidence that soda-rich solutions were present in the late stages of consolidation of the magma, for much of the orthoclase has been irregularly replaced by stringers and patches of albite. This feature is poorly developed in the monzonite.

Calculations show that the quartz-bearing monzonite might have been derived from the monzonite either by the addition of a material approximating a sodic quartz syenite or by the subtraction of a mixture of leucite, anorthite, wollastonite, and iron oxides. The composition of the latter material suggests no common aggregate of minerals. Accordingly, it becomes necessary to search for a more complicated mechanism of differentiation than simple crystal fractionation in order to account for the observed phenomena.

The quartz-bearing monzonite probably came from a monzonitic magma and ultimately from a syenogabbroic magma by a combination of crystal settling, assimilation, or syntexis, and volatile transfer. Pyroxene is much less abundant in the quartz-bearing monzonite than in the monzonite. The process of crystal settling, which resulted in the formation of monzonite from the syenogabbroic magma, probably continued without cessation in the deeper parts of the magma chamber. But crystal fractionation alone does not account for the high silica and alkali content of the quartz-bearing monzonite.

Low quartz biotite granite is the predominant wall rock of the stock, and field observations indicate that large masses of this rock were stoped into the monzonitic magma. Granite blocks, because of their greater density, must have sunk through the liquid monzonite deep into the magma chamber, where they might have been assimilated by the more mafic liquid encountered there.

Bowen (1928), from a consideration of heat supply and the nature of the reaction series, concludes that granite may be assimilated by a more mafic magma and that the process is essentially one of simultaneous solution of felsic materials and crystallization of mafic minerals. Assimilation of granite fragments probably resulted in a rock richer in silica and alkalies than the original monzonite.

Numerous masses of almost completely assimilated granite occur in exposures of the quartz-bearing monzonite and are marked by patches of light-colored coarse-grained quartzose rocks, grading almost imperceptibly into the surrounding rock. The mineral composition of the patches is the

same as that of the granite except that biotite and other mafic minerals are absent.

Irregular albite patches in the quartz-bearing monzonite suggest that soda was introduced. The distribution of the albite is such that its presence cannot be adequately explained by assuming simple exsolution from orthoclase. Accumulation of alkali-rich solutions toward the end of the period of consolidation may have been effective in introducing some soda.

The magnesia content of the quartz-bearing monzonite is less than one per cent, so the subsequent differentiates cannot be reasonably explained by assuming only crystal fractionation involving subtraction of mafic minerals. Syenite follows quartz-bearing monzonite and, according to data obtained from a variation diagram, may have formed either by subtraction of a mixture rich in albite and wollastonite or by the addition of potash feldspar and alkalies. Evidence for both processes is found in field exposures and thin sections. Abundant partly assimilated inclusions suggest extensive contamination by granite. However, andesine and sphene are abundant in thin sections of the quartz-bearing monzonite, and crystals of these minerals may have settled out of molten quartz-bearing monzonite to form a syenitic magma.

Albite is more abundant in the syenite than in the quartz-bearing monzonite. Its distribution suggests replacement, for ragged lenses occupy more than 50 per cent of the orthoclase grains by volume. Albitization of the orthoclase probably was accomplished in part by mobile soda-rich solutions residual from the crystallization of the main mass of the syenite or by solutions rising from deeper parts of the stock.

The mineral composition and texture of granite dikes in the syenite are unusual. Unlike the feldspar in normal granites the orthoclase in the dikes is euhedral to subhedral and appears to have formed by precipitation from a fluid volatile-rich magma.

Increasing albitization of the differentiates with decreasing age implies an increasing accumulation of sodic aqueous solutions with each successive differentiation. Ore deposits related to the stock cut the granite and show evidence of abundant late alkalic solutions.

Von Wolff's (1931) scheme for plotting rock analyses permits an examination of genetic relationships of the members of a differentiation series. By recalculating analyses into ideal aggregates of feldspar and augite and an excess or deficiency of silica, a three component system, which will show natural boundaries of rock families, as well as lines of liquid descent, is obtained. A triangle is used with 200 units on a side as shown in Figures 6 and 7. Figure 7 is an enlargement of the critical portion of Figure 6. On these triangles the point L represents rocks

saturated in quartz; any rocks with a calculated deficiency of silica will, accordingly, lie below the line L-M, and rocks with a computed excess of silica will lie above it. The point M corresponds to a pyroxene fully saturated with silica, and any rocks with a silica deficiency will naturally contain pyroxene with less silica than is necessary for the formation of

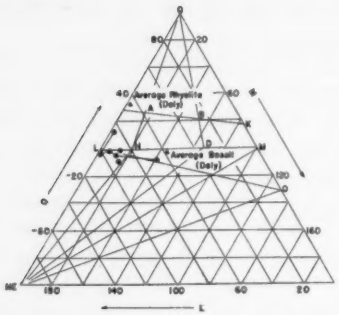

FIGURE 6.—*Von Wolff diagram showing plotted points of rocks in Audubon-Albion stock*

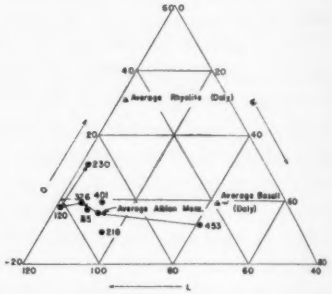

FIGURE 7.—*Enlargement of critical portion of Figure 6*

pyroxene of composition M. The point O indicates the position of olivine on the diagram, NE the position of nepheline, and Q the position of a material consisting of 100 per cent silica. E is the quartz-feldspar eutectic, and K is the cristobalite-clinoenstatite eutectic.

For plotting analyses the sum of computed feldspars is indicated as L or leucocrates, M as the total pyroxene and magnetite or melanocrates, and Q as the excess or deficiency of silica necessary to produce first saturated feldspars, and then pyroxene. The fields of some of the more important rock families are: EKQ, rhyolites, dacites; NE-O-M-H, mafic rocks with predominant soda; L-M-O, rocks with predominating lime including normal basalt; L-M-E-K. intermediate rocks. This diagram cannot be used for plotting modal minerals, unless they closely approach the Von Wolff normative minerals. However, the resemblance between normative and modal minerals is very close in many rocks, and under such circumstances the diagram is especially valuable.

The diagram is useful in the study of differentiation by crystal fractionation. If the composition of a magma and the crystals removed from it are known, the change of composition of the residual magma can be shown by drawing a straight line through the plotted points of the original magma and the substracted crystals. The composition of the melt will change along the line in a direction opposite to that of the fractionated

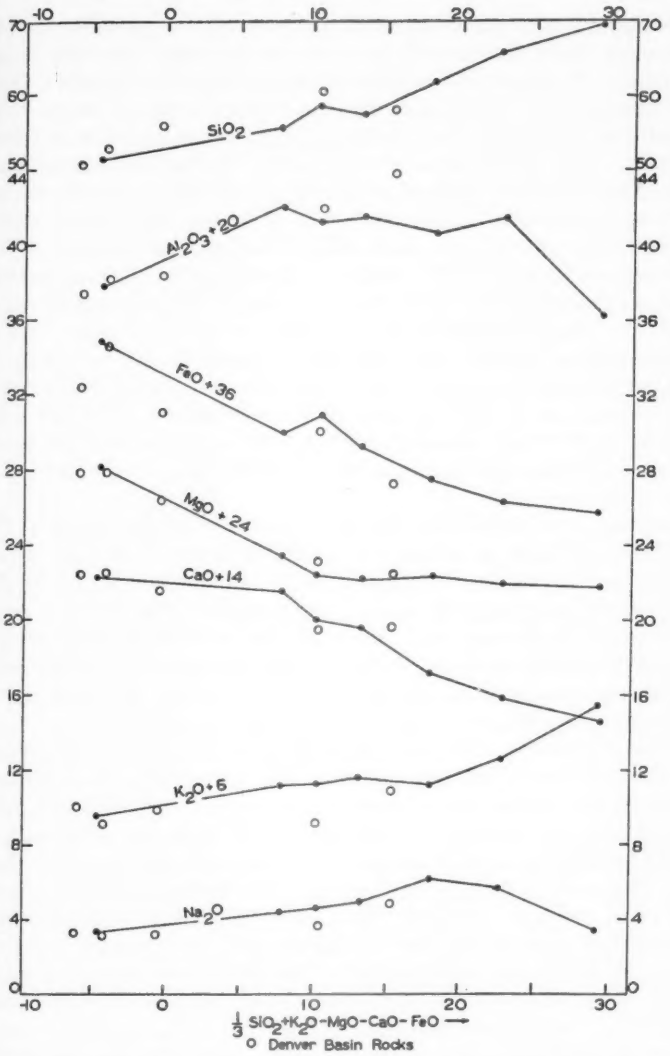

FIGURE 8.—*Larsen diagram of Audubon-Albion rocks and rocks of Denver Basin*

crystals. If two or more minerals are removed by crystallization, the course of liquid descent will be a curved line whose direction is controlled by the nature and relative amounts of separated crystals. From the diagram (Fig. 7) it is seen that the average monzonite might be derived from the syenogabbro (specimen 453) by the separation of a mafic material approaching augite in composition. Beyond the monzonite the line of liquid descent changes its course, and in order to obtain the next differentiate—quartz-bearing monzonite (specimen 326)—it is necessary to postulate removal of a mafic silica-undersaturated material of very unusual composition or the addition of a feldspathic siliceous material. The quartz-bearing monzonite probably resulted from concurrent assimilation of granite inclusions and separation of mafic minerals. Further assimilation of granite might form syenite (specimen 120) from a quartz-bearing monzonitic magma; in any event, the variation diagram indicates little separation of mafic minerals to produce syenite. The last differentiate in the stock—granite (specimen 230)—was formed from syenite by special processes, as indicated by the abrupt change in the trend of the line of liquid descent.

Another type of variation diagram is shown in Figures 8 and 9. The diagrams are based on a scheme proposed by Larsen (1938) and provide means for showing the variation of rocks between basalt and rhyolite. The plotted analyses of the rocks of the Audubon-Albion stock show differentiation progressing from a rock similar to basalt in composition to a rock not unlike rhyolite. In general, the differentiation series shows a steady decrease in lime, magnesia, and iron oxides, and a concurrent uniform increase in potash and silica with decreasing age.

Figure 8 shows a Larsen variation diagram with the plotted analyses of the Audubon-Albion stock (dots) and, for comparison, analyses of rocks of the nearby Denver Basin (open circles). Figure 9 is a generalized diagram in which the Audubon-Albion rocks are compared with rocks from the San Juan Mountains in Colorado and the Crazy Mountains and Highwood Mountains in Montana. The Audubon-Albion rocks are lower in magnesia and iron oxide than rocks from other provinces. Potash and soda are high but not so high as in the Highwood rocks.

In conclusion, the differentiates of the Audubon-Albion stock appear to have been formed in the early stages of the intrusion by separation of crystals of mafic minerals and calcic plagioclase and in the later stages by more complex processes involving not only crystal fractionation but assimilation and transfer of alkalic minerals and silica by highly mobile fluids.

CONTACT METAMORPHISM

Contact metamorphism in the rocks adjacent to the stock is not great. Positive evidence that the intrusion affected its walls is found only in

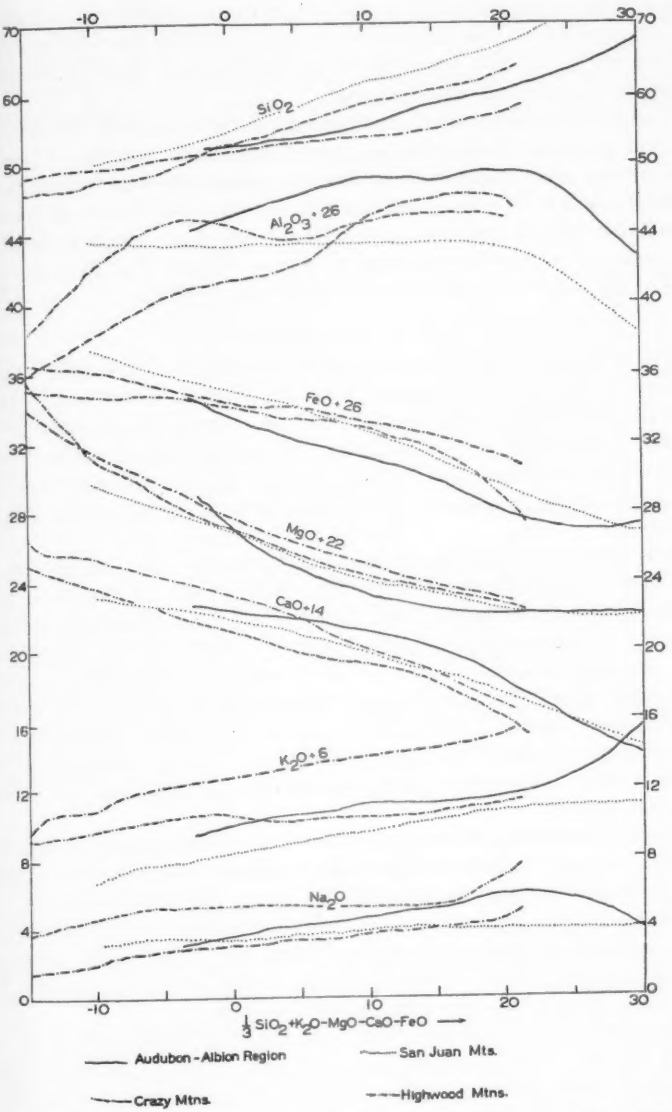

FIGURE 9.—*Generalized Larsen diagram of rocks in Audubon-Albion stock and three other igneous provinces*

veins of magnetite and joints coated with pyroxene, amphibole, and sphene in a narrow zone a few hundred feet wide surrounding the western border of the stock. In most places adjacent pre-Cambrian rocks do not differ essentially from rocks at a distance.

The greatest changes resulting from contact metamorphism are observed in roof pendants of amphibolite exposed near Lake Isabelle. Chemical and petrographic studies indicate that the amphibolite was formed by the vigorous action of silica-undersaturated solutions on masses of foliated pre-Cambrian pyroxene rock and quartzite. Pyroxene gneiss is abundant immediately west of the amphibolite masses. Figure 2 of Plate 4 shows one of the larger amphibolite masses near Lake Isabelle.

MODE OF EMPLACEMENT

The composite Audubon-Albion intrusion consists of five different and separately intruded rock types. The map (Fig. 2) of the stock shows partial development in the northern half of a crude, unsymmetrical concentric arrangement of the intrusions. In contrast dikelike bodies characterize the southern half. These appear to have followed contacts within the stock or contacts of the stock with the pre-Cambrian granite and gneiss.

The intrusive mechanism probably was complex. The distribution of the rocks in the northern half of the stock suggests the ring-dike structures of Scotland and elsewhere. Flow layering and linear structures indicate that the internal contacts dip steeply outward, so that in three dimensions the structure consists essentially of a series of discontinuous nested truncated cones. The structure and sequence of intrusions might well be explained by assuming intermittent sinking of a large conical block in the manner suggested by Richey (1928) and further elaborated on by Daly (1933). The name Daly suggests for the mechanism is "ring-fracture stoping".

Breccia zones of pre-Cambrian rocks between the contacts of the successive Tertiary intrusions add to the argument. These zones resemble the so-called "screens" that characterize many ring-dike complexes.

In the southern half of the stock intruding magmas either have forced their way into contact zones or stoped their way into zones of weakness along contacts. There is little evidence of the former mechanism, for the stock sharply truncates structures in its walls.

The stock locally contains abundant angular inclusions (Pl. 3, fig. 1; Pl. 4, fig. 1). Most of the inclusions are only a few inches in diameter, but many attain diameters of 30 or 40 feet. Probably, then, the intrusives made room for themselves in part by "piece-meal stoping."

Balk (1937) recently has suggested that many igneous masses that

appear to have been emplaced by stoping actually may have forced their way into the earth's crust, but the evidence for this process may be found only in structures at some distance from the intrusive. Distant structures that might have formed by the pushing aside of the walls of the Audubon-

TABLE 2.—*Chemical Analyses*

	I	II	III	IV	V	VII	VII
SiO_2	52.59	49.69	50.37	59.26	59.29	56.62	62.27
TiO_2	0.84	0.85	1.53	.....	.....	0.92	0.12
Al_2O_3	17.91	18.06	17.38	23.63	21.27	21.39	21.70
Fe_2O_3	3.81	2.64	1.91	0.30	3.33	1.18	0.61
FeO	5.18	6.19	6.06	0.57	1.04	2.18	0.41
MgO	4.11	5.73	6.16	0.31	1.12	0.29	0.04
CaO	7.24	8.24	8.25	5.93	5.25	5.64	1.94
Na_2O	2.94	2.99	3.16	4.94	3.39	4.91	5.84
K_2O	3.83	3.90	3.60	4.78	3.00	5.63	6.86
H_2O	1.24	0.91	0.57	0.74	1.63	0.56	0.53
P_2O_5	0.14	0.81	0.54	.....	0.23	0.36	.....
MnO	Trace	0.13	0.13	.....	0.21	0.06	.....
	99.83	100.14	99.66	100.46	99.76	99.74	100.32

I. Basalt. Capping sheet, Table Mountain, Golden, Colorado. United States Geol. Survey, Monograph 27, p. 306. Analyst: W. F. Hillebrand.
II. Basalt. Early basal sheet, Table Mountain, Golden, Colorado. United States Geol. Survey, Monograph 27, p. 308. Analyst: L. G. Eakins.
III. Syenogabbro. Lake Isabelle, Boulder County, Colorado. Analyst: E. E. Wahlstrom.
IV. Andesite pebble. Denver formation, United States Geol. Survey, Bull. 148, p. 159. Analyst: L. G. Eakins.
V. Same as IV.
VI. Monzonite. Lake Albion. Analyst: E. E. Wahlstrom.
VII. Syenite. Lake Albion. Analyst: E. E. Wahlstrom.

Albion stock are absent. For a radius of several miles around the stock the pre-Cambrian rocks are only slightly disturbed.

REGIONAL RELATIONSHIPS AND AGE

Lovering and Goddard (1938a) state that the Audubon-Albion stock probably filled the throat of a volcano that supplied lava pebbles to the Denver formation of late Cretaceous age.[1] Data assembled by these writers indicate that the intrusion of the Audubon-Albion stock is one of the latest events in the Cretaceous period and foreshadowed the long and complex Tertiary igneous history of the Front Range. Additional data collected by the present writer lead to the conclusion that igneous activity

[1] These writers (1938b) also have suggested that the Denver formation is possibly Eocene. The present writer has discussed its age with many geologists. The consensus of opinion is that the formation is Eocene or Paleocene.

in the stock probably began in early Eocene or Paleocene and lasted well into the Tertiary.

Lovering and Goddard (1938a) determine the time of intrusion by correlating analyses of andesitic pebbles in the Denver formation with analyses of monzonite and syenite from the Audubon-Albion stock. The correlation is questionable. In Table 2 analyses used by the above writers and analyses used by the present writer are tabulated. Analyses IV and V were made on andesitic pebbles from the Denver formation collected at some distance below the capping basalt flows at Table Mountain near Golden, Colorado. Analyses VI and VII were made by the present writer and, because of certain similarities to analyses IV and V, were used by Lovering and Goddard to date the intrusion of the Audubon-Albion stock. The analyses are all low in magnesia and high in alumina. Analyses I and II were made on samples from the capping and basal basalt flows on Table Mountain; analysis III was made on a sample of syenogabbro from Lake Isabelle. It is evident that the similarities among I, II, and III are much more obvious than those among analyses IV, V, VI, and VII. However, all the rocks probably are closely related, for analyses of rocks from the Denver Basin plotted on a variation diagram of the Audubon-Albion rocks show fair agreement (Fig. 8). The agreement is excellent near the low silica end of the curves; and, when it is considered that the andesite pebbles from the gravels in the Denver formation are much altered, the agreement toward the high silica end is within reasonable limits.

Chemical similarities of the basalt and syenogabbro lead to the conclusion that the oldest rock in the Audubon-Albion stock is more closely related to basalt flows in the upper part of the Denver formation than to andesitic materials in its lower part. But the general similarity of all the analyses may be used to conclude safely that the igneous activity in the two regions probably was in part contemporaneous.

Of doubtful value is physiographic evidence indicating that the stock is very early Tertiary. The so-called "Flat-top" peneplain truncates the intrusion, and several writers, among whom are Little (1925), Lee (1923), Mather (1925), and van Tuyl and Lovering (1935), have stated that the peneplain is, at least partly, Eocene.

SUMMARY OF CONCLUSIONS

The composite Audubon-Albion stock in Boulder County, Colorado, consists largely of monzonite and crosscuts pre-Cambrian granite, gneiss, and schist.

The oldest rock in the stock—syenogabbro—is believed to be a chilled phase of the parent magma from which the other types of rock were de-

rived. Monzonite probably formed from the syenogabbroic magma by a process of crystal fractionation, but younger differentiates, including quartz-bearing monzonite, syenite, and granite, resulted from more complex processes involving not only crystal fractionation but also assimilation or syntexis and transfer of materials by mobile, fluid-rich, alkalic solutions. The sequence of intrusions is syenogabbro (oldest), monzonite, quartz-bearing monzonite, syenite, and granite.

Age relations, the distribution of the various rock types, and internal structures suggest that the stock was in part emplaced by repeated subsidence of a large conical block and in part by the rise of stoping magmas into zones of weakness along contacts within the stock or contacts between the stock and older rocks. Evidence is lacking to indicate that the walls of the stock were forcibly separated by advancing magma.

Chemical analyses of the syenogabbro show an exceptionally high content of potash and are nearly identical with analyses of the basalt flows interbedded with the sediments of the Denver formation near Golden, Colorado. Correlation establishes the probable age of the oldest rock in the Audubon-Albion stock as early Eocene or Paleocene. The exact age of the younger differentiates is uncertain.

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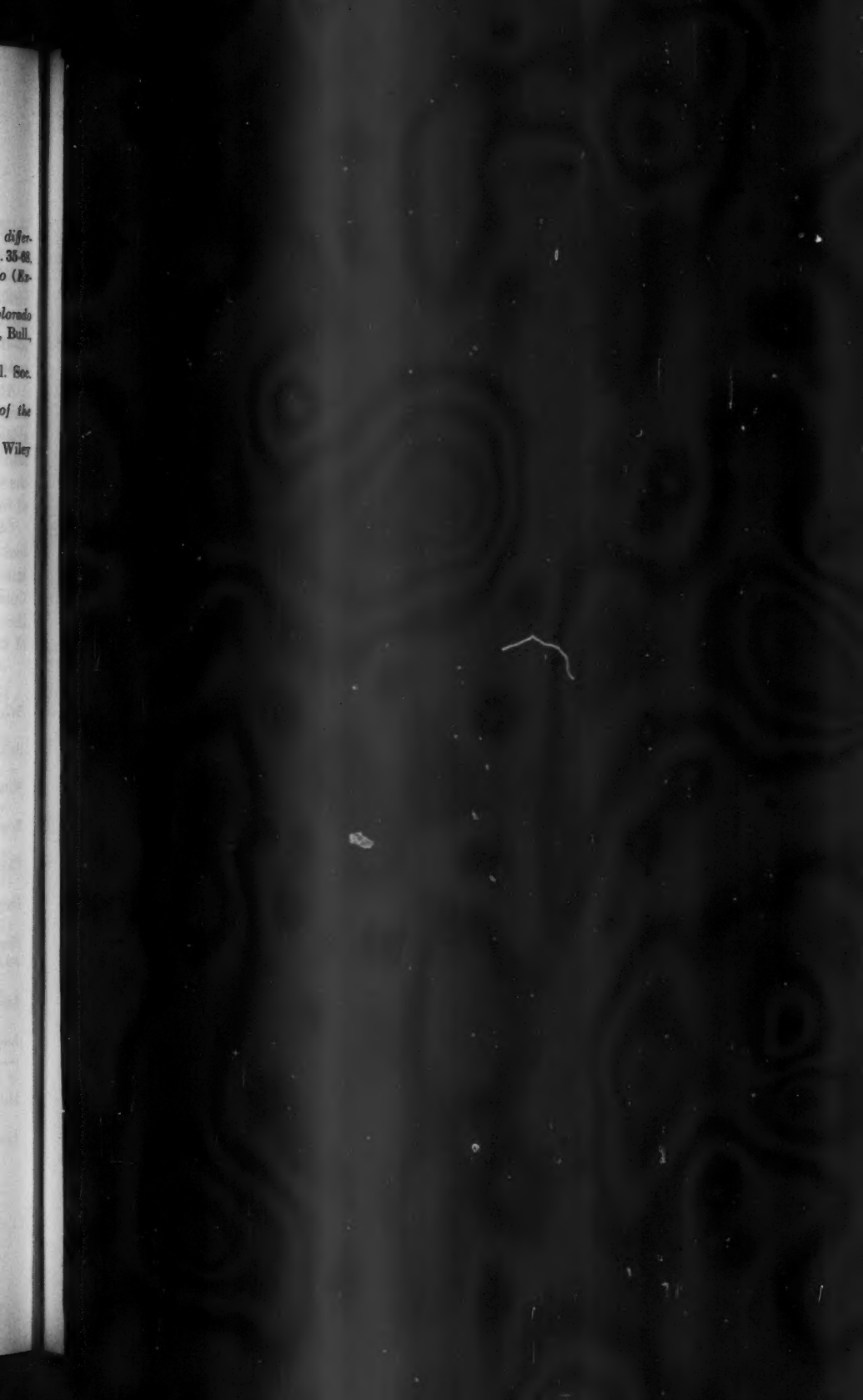

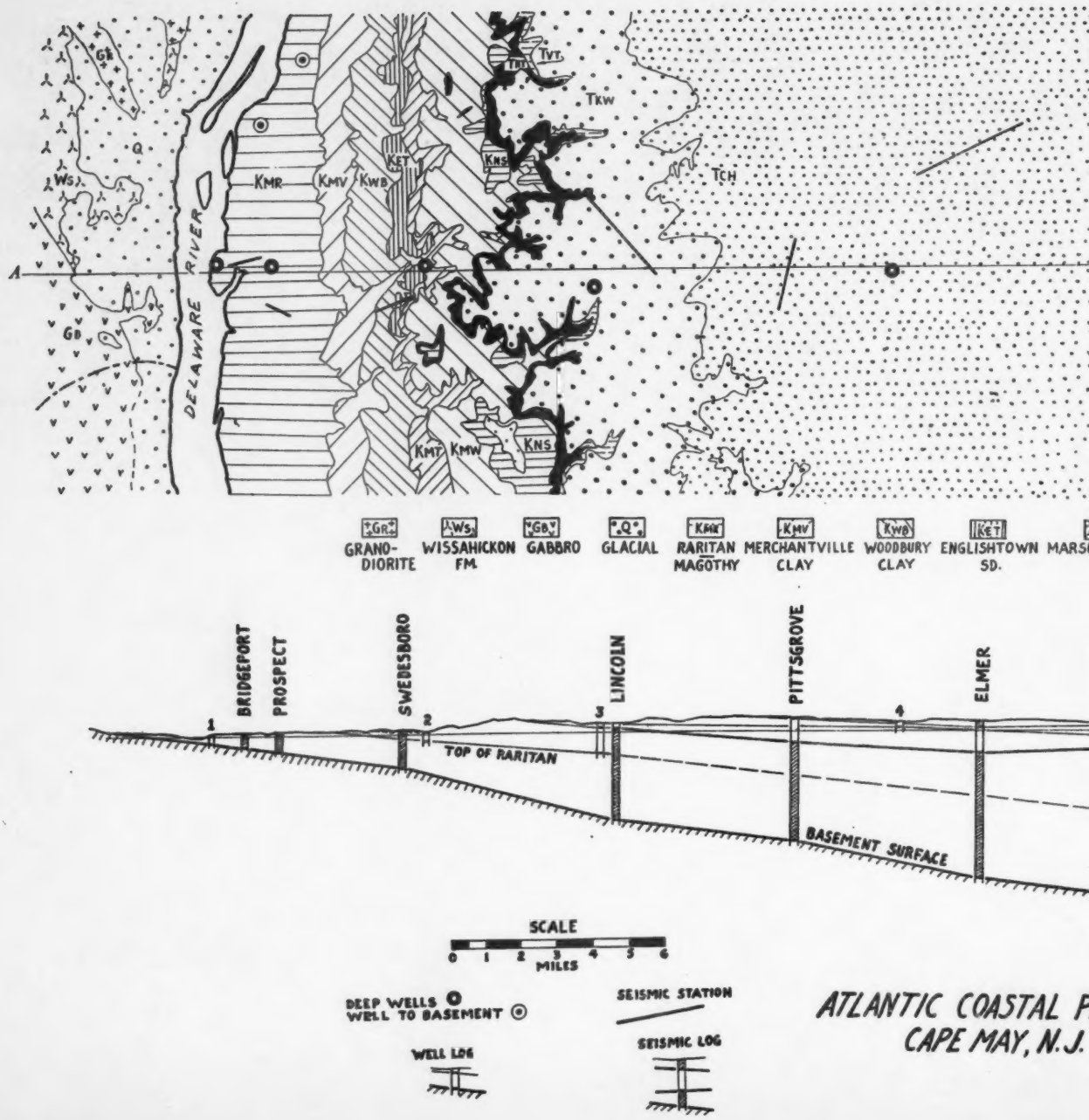

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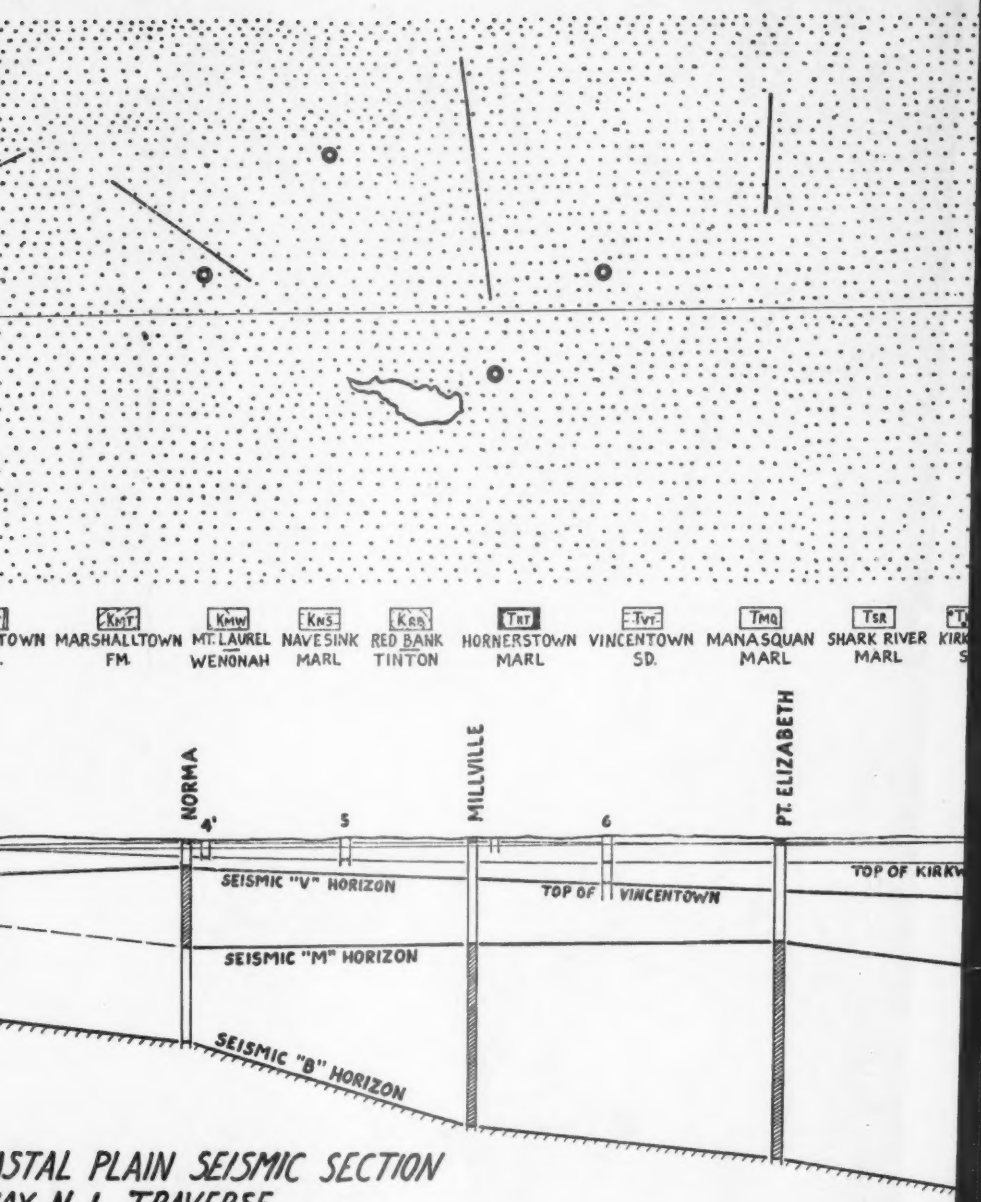

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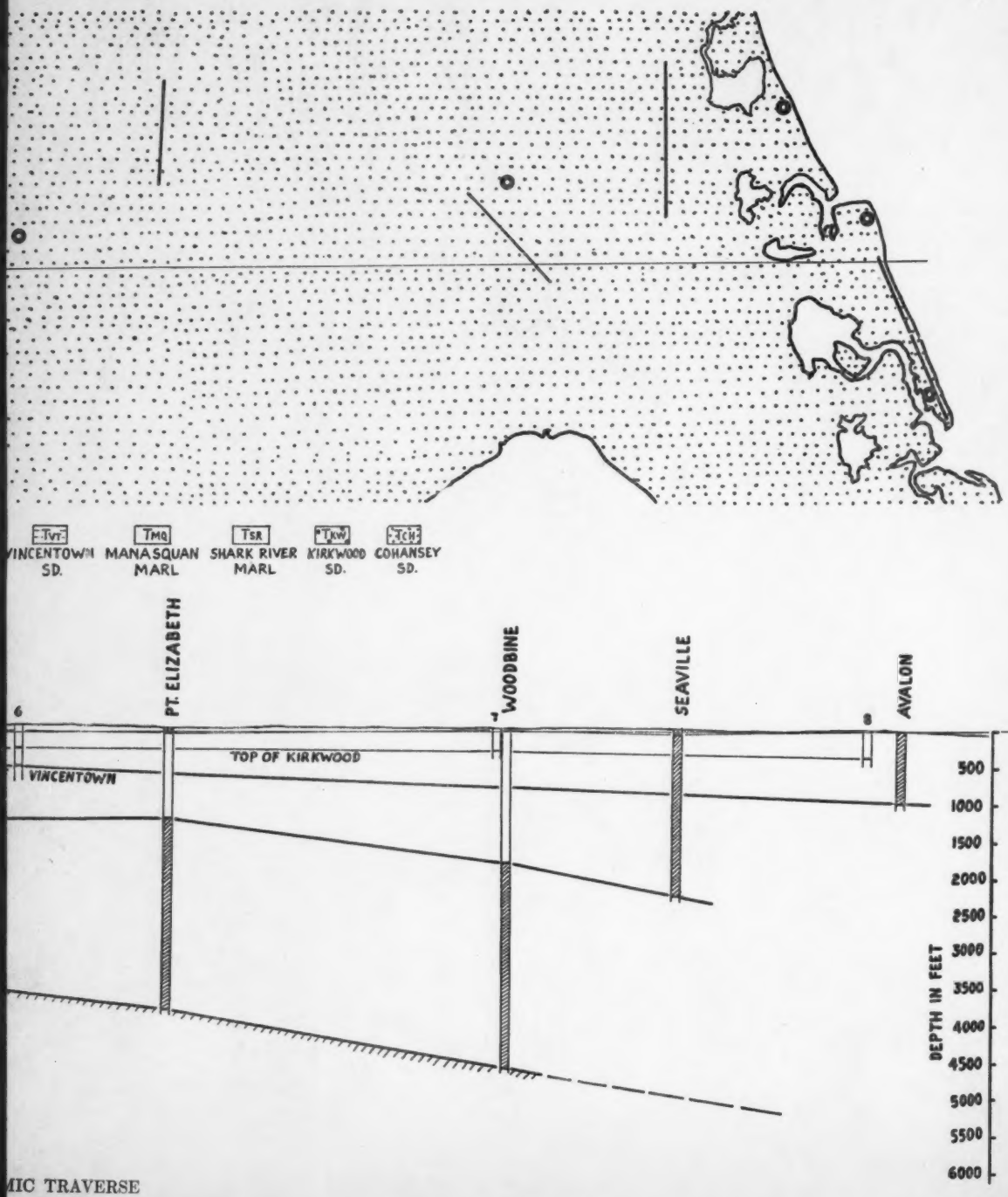

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GEOPHYSICAL INVESTIGATIONS IN THE EMERGED AND SUBMERGED ATLANTIC COASTAL PLAIN

PART IV. CAPE MAY, NEW JERSEY, SECTION

BY MAURICE EWING, GEORGE P. WOOLLARD, AND A. C. VINE

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ABSTRACT

Seismic refraction measurements across the New Jersey Coastal Plain, from Bridgeport on Delaware Bay to Avalon on the Atlantic coast, have been made as a continuation of similar measurements of the depth of the crystalline basement made in 1935 at Cape Henry, Virginia, and Woods Hole, Massachusetts, and in 1937 at Barnegat Bay, New Jersey. The standard technique for seismic refraction measurements was followed. Three seismic discontinuities were determined in addition to the one associated with the bottom of the surface gravels covering this part of New Jersey. These discontinuities appear to have the same geologic equivalents as were noted on the Barnegat Bay seismic section. The observed depths are in agreement with the available well data, and the seismic velocity variations noted for the basement discontinuity indicate similar changes in basement lithology as those found along the Barnegat Bay seismic section.

FOREWORD

In order to avoid the repetition of data and diagrams the writers present the following seismic investigations of the subsurface geological structure in the Coastal Plain of New Jersey as one composite paper, covering both the seismic investigation and the geological significance of the results.

The geological interpretations are the work of Woollard, and the least square solution used in reducing the data was devised by Ewing. All the writers collaborated in conducting the field work, in reducing the data, and in preparing the report.

INTRODUCTION

The writers present a fourth seismic refraction profile in the study of the fundamental structural conditions along the Continental Shelf. The section presented begins near the Bridgeport Ferry House on Delaware Bay and ends at Avalon on the Atlantic shore. Twelve land seismic stations were established at intervals of about 6 miles. Reverse profiles were run at only seven stations for local cultural conditions and natural obstacles prevented use of the reverse profile technique at the other five stations. At all the stations slight slopes are indicated for both the basement crystallines and the intermediate stratigraphic horizons detected.

For the five stations which were not reversed, the method of conducting the seismic survey was the same as that described for land stations in Part I (Ewing *et al.*, 1936, p. 759). Where reverse profiles were run, the method was as described in Part III (Ewing *et al.*, 1939, p. 262). At

all stations the time data were least squared so that the best possible velocity graph could be obtained. The method of least squares used will be discussed in detail.

As on the Barnegat Bay section, certain seismic horizons in the overlying sediments could be traced accurately and appear to have the same geologic equivalents. This is indicated by the similarity in the observed velocities and the depth correlation obtained from well data.

The relation observed between the seismic velocities for the basement crystalline rocks and the gravity anomalies in Virginia in Part III (Ewing *et al.*, 1939, p. 283) also appears to hold for both the Barnegat Bay and the Cape May sections in New Jersey, and it now appears possible, within limits, to differentiate the nature of the basement rocks by these geophysical measurements.

ACKNOWLEDGMENTS

The writers acknowledge the aid extended by the New Jersey Division of Geology and Topography, through Meredith E. Johnson, State Geologist. To this branch of the New Jersey Department of Conservation and Development they are indebted not only for scientific assistance but also for financial help in carrying out the project.

For assistance in carrying out the field work, the writers wish to acknowledge the help extended by J. L. Worzel, R. J. McCurdy, and N. H. Webster of Lehigh University.

To The Geological Society of America the writers are indebted for the use of field equipment constructed from funds allocated for carrying out seismic investigations in the oceanic areas.

Ewing and Woollard are indebted to the Guggenheim Foundation and to the National Research Council, respectively, for fellowships covering the period of these researches.

LOCATION OF SECTION

The line of the section was chosen to be essentially perpendicular to the strike of the structures being investigated and also located so as to give the maximum amount of geologic information in an area where there are practically no deep subsurface data. It traverses the southern part of New Jersey, an area about 60 miles wide containing few deep wells. In laying out the station profiles, an effort was made to have their lines as nearly perpendicular to the line of the section as possible as this was believed to be the direction of least slope in the basement. However, natural obstacles and cultural conditions were more frequently the deciding factors in establishing these station profiles. Figure 1 shows

the relative position of the four seismic sections established to date in this study of the structural conditions along the continental border. Plate 1 indicates the location of the Cape May line in detail with the 12 station profiles.

The seismic stations established and the distances between them are as follows:

Bridgeport			5.0
	1.2	Norma	
Prospect			7.0
	3.5	Millville	
Swedesboro			7.3
	6.0	Port Elizabeth	
Lincoln			9.3
	5.0	Woodbine	
Pittsgrove			4.7
	5.3	Seaville	
Elmer			6.3
	5.0	Avalon	

METHOD OF INVESTIGATION

The standard method of refraction seismic measurements as described in Part I (Ewing *et al.*, 1936, p. 759) was employed, with a few instrumental changes as indicated in Part III (Ewing *et al.*, 1939, p. 262).

The records for all stations were interpreted by the method described on page 262 of Part III, except for Swedesboro, Port Elizabeth, Woodbine, and Avalon, which were analyzed by the method set forth in Part I (Ewing *et al.*, 1936, p. 763), as the slope indicated was too small to justify more elaborate calculations.

METHOD OF LEAST SQUARES

In Part III, page 263-264, the basic relations which must be satisfied by a group of overlapping profiles lying on a common line are explained. They are: (1) The "reverse points" of any oppositely directed overlapping pair of profiles must be equal, and (2) The "time-intercept points" for each layer for all profiles of the group must be collinear. A solution, based on the method of least squares, has been derived for determining a self-consistent set of travel-time curves to represent the travel-time data.

Figure 2 shows a set of profiles, A_1, A_2 . . . extending in one direction, and B_1, B_2 . . . extending in the opposite direction along the same line. Ordinates (t) represent travel times, and abscissas (x) represent the positions of the shot points and recording points between which travel times were measured. Thus t_1 represents the travel time from x to

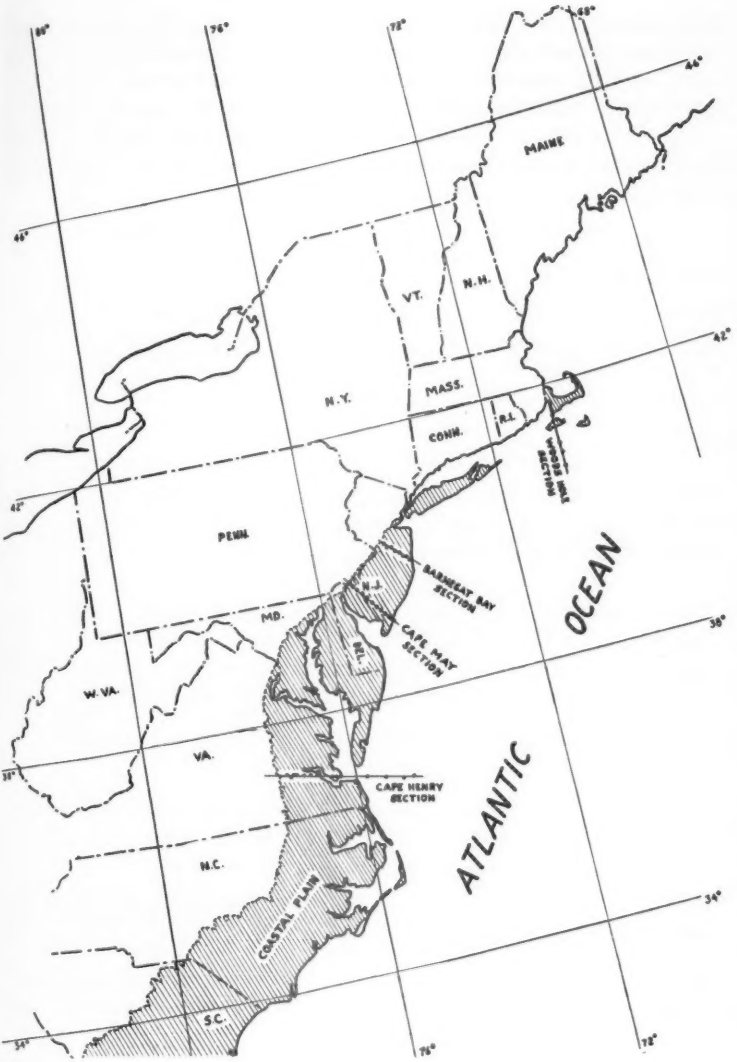


FIGURE 1.—*Index map showing location of geophysical traverses across Atlantic Coastal Plain*

B_1, t_2 that from x to B_2, t_3 that from x to A_2, etc. The time-intercept points for all profiles lie on CD, and T represents the ordinate of this line at $x = O$. The slopes (or reciprocal apparent velocities) are m and n respectively, as indicated in Figure 2. In general m is positive, n negative.

Since E and F are reverse points and therefore have equal ordinates, the difference in ordinate between points C and D is $(m + n)B_1$, and the slope of CD is $m + n$. It follows that

$$T_{A1} = T + (m + n)A_1,\ T_{A2} = T + (m + n)A_2, \text{ and so on.}$$

Consider profile A_1.

$$t = T_{A1} + m(x - A_1) = T + (m + n)A_1 + m(x - A_1) = T + mx + nA_1.$$

Similarly, for profiles A_2, A_3, B_1, B_2, etc., the equations are:

$$\begin{aligned} t &= T + mx + nA_2 \\ t &= T + mx + nA_3 \\ t &= T + mB_1 + nx \\ t &= T + mB_2 + nx \end{aligned}$$

Let x_i, t_i be the coordinates of the observed travel time values. Then the deviations of the observed points from the theoretical lines are expressed by terms of the form

$$\left.\begin{aligned} \epsilon_{11} &= T + mx_1 + nA_1 - t_1 \\ \epsilon_{12} &= T + mx_2 + nA_1 - t_2 \end{aligned}\right\} \text{ for } A_1$$

$$\left.\begin{aligned} \epsilon_{21} &= T + mx_1 + nA_2 - t_1 \\ \epsilon_{22} &= T + mx_2 + nA_2 - t_2 \end{aligned}\right\} \text{ for } A_2$$

$$\left.\begin{aligned} \epsilon_{11} &= T + mB_1 + nx_1 - t_1 \\ \epsilon_{12} &= T + mB_1 + nx_2 - t_2 \end{aligned}\right\} \text{ for } B_1$$

etc.

The following normal equations are derived from the above equations of observations by the standard method.

$$(1)\quad m(\Sigma_A x^2 + \Sigma_B N_B B^2) + n(\Sigma_A Ax + \Sigma_B Bx) + T(\Sigma_A x + \Sigma_B N_B B) = \Sigma_A tx + \Sigma_B tB$$
$$(2)\quad m(\Sigma_A Ax + \Sigma_B Bx) + n(\Sigma_A N_A A^2 + \Sigma_B x^2) + T(\Sigma_A N_A A + \Sigma_B x) = \Sigma_A tA + \Sigma_B tx$$
$$(3)\quad m(\Sigma_A x + \Sigma_B N_B B) + n(\Sigma_A N_A A + \Sigma_B x) + TN = \Sigma_A t + \Sigma_B t.$$

Each summation is extended over all observed points on either the A or B profile, as indicated by the subscript. N_A and N_B represent the total number of observed points on each A and B profile respectively. N represents total number of observed points on all profiles. From these

equations the values m, n, T are computed. These equations were used for the computation of the constants m, n, T for each station included in this report. Removal of personal judgment in drawing the travel-time curves, rather than an increase in precision of the final result, is the primary reason for using a least square solution.

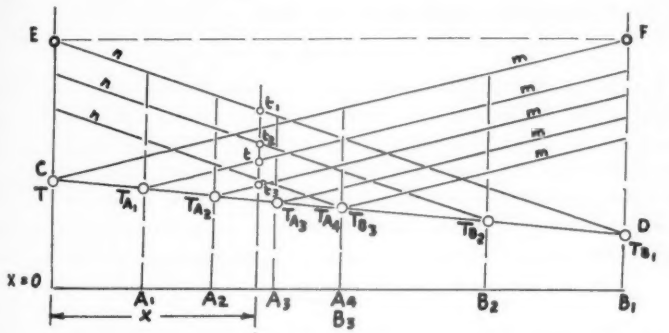


FIGURE 2.—*Time-distance graph showing values and relations used in least square solution*

As stated in Part III (page 263) this system of calculation yields a plane surface for the interface between each pair of seismic layers. Where the actual surface deviates slightly from a plane, the solution gives the plane which best approximates the actual surface.

TRAVEL-TIME DATA AND GEOPHYSICAL RESULTS

GENERAL STATEMENT

In this section all of the travel-time data are presented in the form of graphs (Pls. 2-4). Along with these data are shown the respective station maps on which the location of the various shots and geophone spreads used for picking up the seismic waves are indicated.

The equation of each straight line in the travel-time graphs was determined by the method of least squares, and the apparent velocity so obtained is written beside the line. These data together with the depths and velocities calculated from them are summarized in Tables 1 and 2.

PROSPECT

The Prospect station is on the Delaware Bay end of the section line. The results obtained at Prospect are extremely good, as indicated by the travel-time plot. No intermediate seismic horizon in the overlying sediments was recorded, and this is what would be expected in the light

of the known geology and the past seismic experience of the writers in New Jersey. The surface sediments are predominantly Recent and overlie a thin layer of the Raritan formation, which is the formation just below the lowest seismic stratigraphic discontinuity recorded so far in these investigations. The depth of the basement discontinuity checks with near-by well data, and the basement is characterized by a velocity of 17,560 ft./sec. and a slight dip to the south.

SWEDESBORO

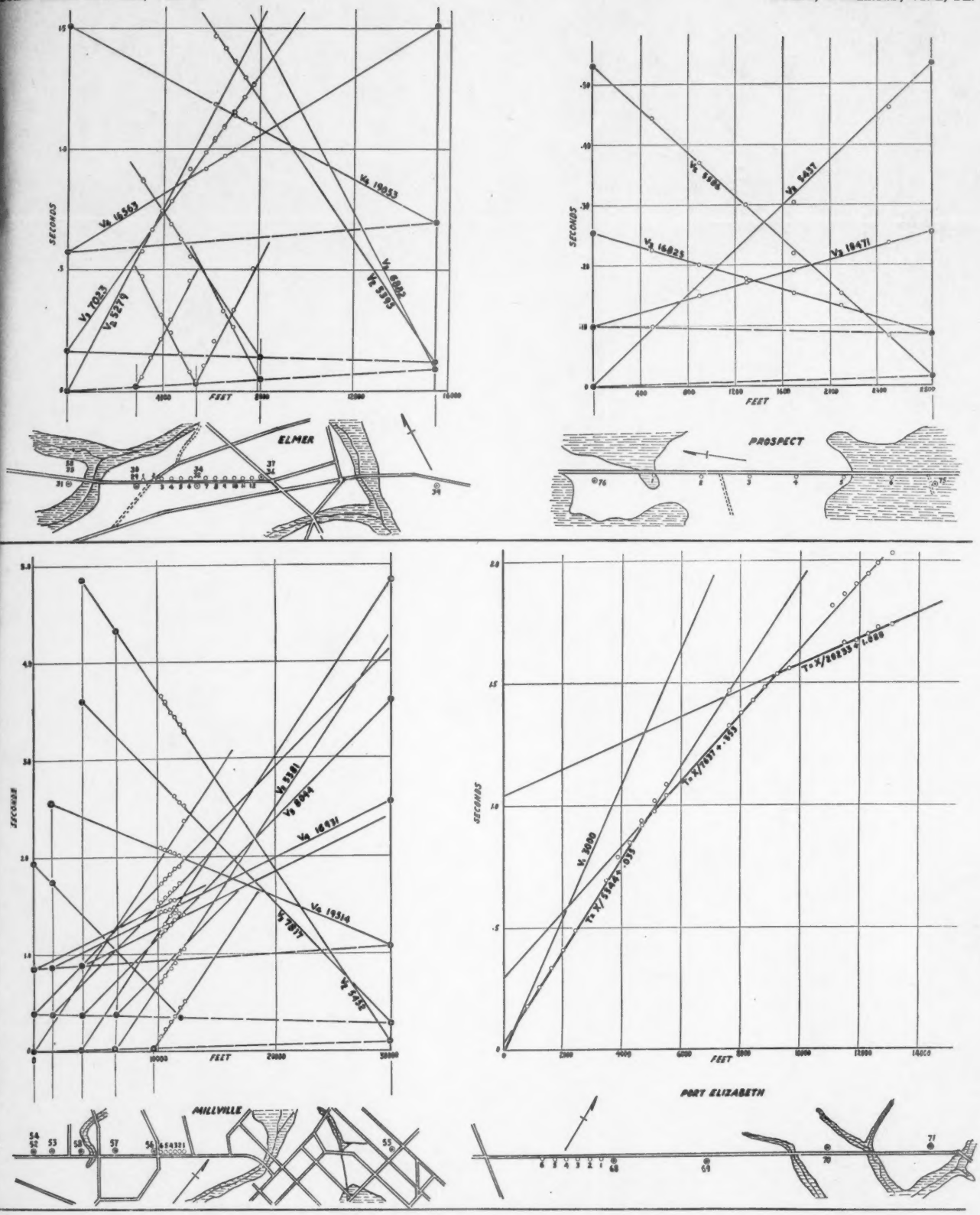
The Swedesboro station graph shows a positive intercept for the low velocity line indicating the presence of the seismic *"M"* discontinuity which occurs at about the unconformity between the Raritan and Magothy formations. For purposes of calculation, the upper layer was assumed to have the same velocity noted for the surface layer at the other seismic stations. This was 3000 ft./sec. and appears to be peculiar to the Cape May gravels which cover nearly all of southern New Jersey.

LINCOLN

At Lincoln geologic conditions were more like Prospect than at most of the other stations in that the Cape May gravels were absent over part of the profile. The depth of the first seismic discontinuity was 67 ft. and from the geological section (Pl. 1) it is seen that this corresponds in position to the top of the Vincentown formation as determined from the log of an adjacent well. The velocity found for this horizon agrees with that of the *"V"* horizon observed on the Barnegat Bay section, which is also associated with the Vincentown formation. The basement discontinuity indicates a slight slope to the north, and the basement rocks apparently differ from those lying between Bridgeport and this point in that they are characterized by a velocity of 19,582 ft./sec. in contrast to the average value of about 17,300 ft./sec. observed for the stations up to this point on the section.

PITTSGROVE

At Pittsgrove, as at Lincoln, the Cape May gravels are present over only a portion of the station line. As a result of this condition, plus the ironing out effect of the method of least squares, a slight negative intercept is indicated for the discontinuity associated with the bottom of the surface gravels. The effect of this discrepancy with theory is, however, so slight that it is regarded as negligible. The seismic discontinuity detected in the overlying sediments corresponds to the *"V"* horizon as observed at Lincoln. The basement rock character has changed so that its velocity is 17,560 ft./sec., and there is a well-defined slope to the west.



SEISMIC STATION PLOTS AND TRAVEL-TIME GRAPHS

ELMER

The Elmer travel-time graph shows three velocity lines in addition to the surface velocity. The *"V"* horizon is again present, and the basement has a normal velocity of 17,690 ft./sec. Again a slight slope to the west is indicated. These basement slope indications are apparently local, as the section line graph shows a regional slope to the east.

NORMA

At Norma three velocity lines are indicated in addition to the surface velocities. When plotted they appear to define the seismic *"V"* horizon, the *"M"* horizon, and the basement. This is the only station where all horizons were indicated. At the others only one of the intermediate stratigraphic horizons was detected, in addition to the surface velocity layer and the basement. They are probably all present throughout the section, but at the various stations one apparently locally masks or incorporates the other. For the basement, a slight slope to the north is indicated, and the basement rock has changed sufficiently to raise the velocity to 18,040 ft./sec.

MILLVILLE

At Millville the *"V"* horizon appears to be masked, and only the *"M"* horizon is indicated. As at Norma, the basement velocity is a little over 18,000 ft./sec., and the basement has a slight slope to the southwest.

PORT ELIZABETH

The Port Elizabeth graph shows no slope condition for the basement, but the indicated velocity is higher than that observed previously—20,233 ft./sec. The value was determined by least squares and is probably very close to the true value. On the basis of relative geographic distribution of velocities, it corresponds in position very closely to the Charleston Springs value on the Barnegat Bay section, which had a true velocity determination of 21,060 ft./sec. As at Millville, the intermediate stratigraphic horizon indicated is the *"M"* horizon.

WOODBINE

This is the last station on the section at which a measurement of depth to basement was obtained. The observed velocity for the basement rocks is 16,639 ft./sec., which fits in with the distribution of velocities encountered on the Barnegat Bay Section. As at the two previous stations, only the *"M"* horizon is indicated in the overlying sediments.

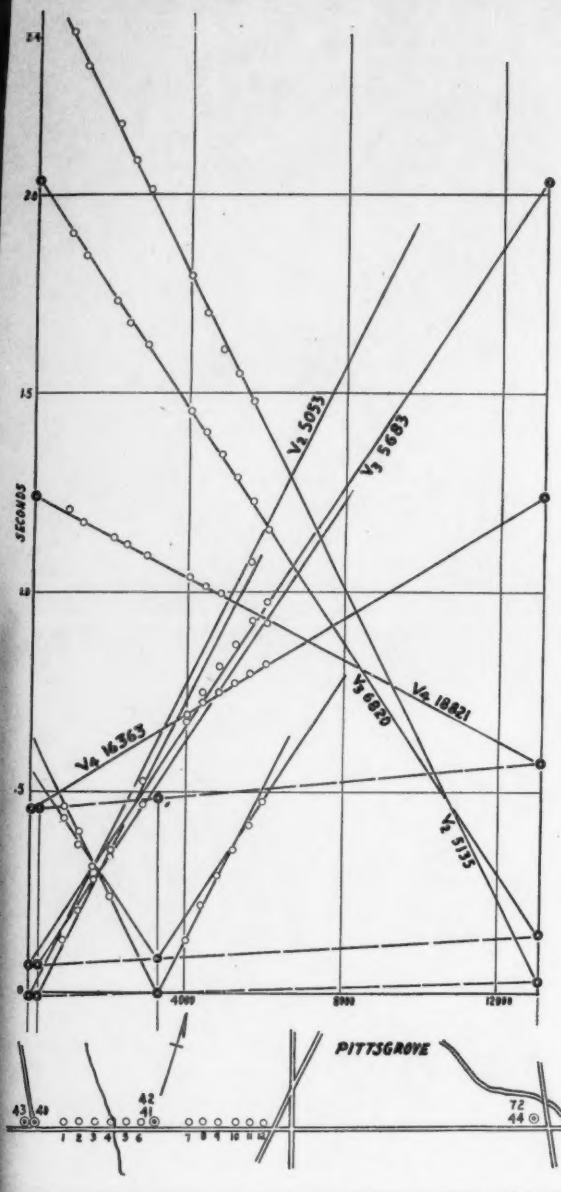
SEAVILLE

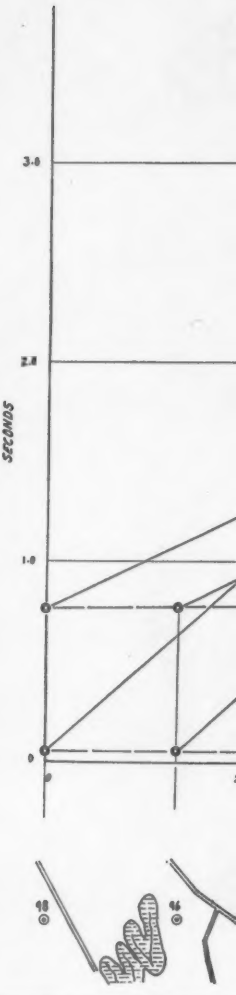
The Seaville station was established between Woodbine and Avalon, as it was impossible to get a record from the basement at the latter

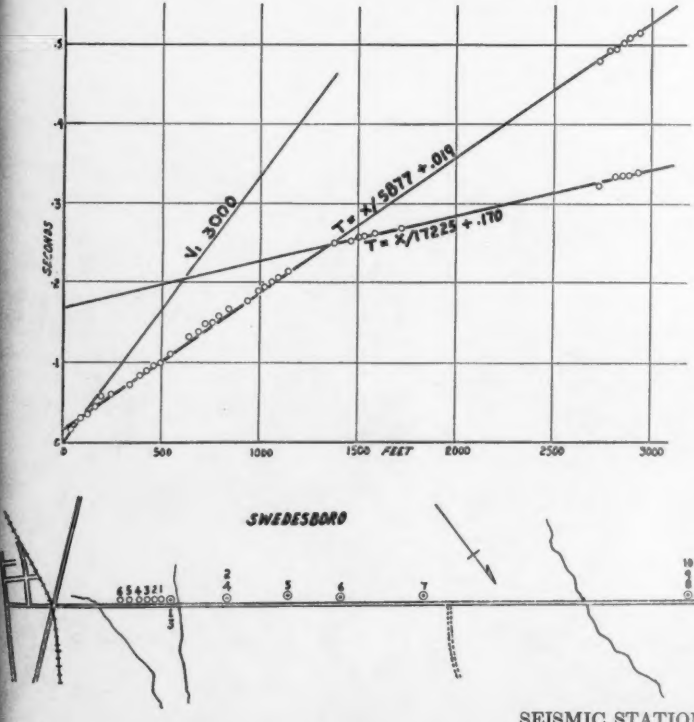
TABLE 1.—*Observed velocities and basement slope*

Prospect:			
V_1 = 3000	V_{2b} = 5586	V_1 = 3000	T_{2a} = .0018
V_{2a} = 5437	V_{3b} = 16824	V_2 = 5510	T_{3a} = .0995
V_{3a} = 18471		V_3 = 17598	w = 1° 50.43′
Swedesboro:			
V_1 = 3000		V_1 = 3000	T_2 = .019
V_{2a} = 5877		V_2 = 5877	T_3 = .170
V_{3a} = 17225		V_3 = 17225	
Pittsgrove:			
V_{2a} = 5053	V_{2a} = 5135	V_1 = 3000	T_2 = .001
V_{3a} = 5683	V_{3b} = 6820	V_2 = 5100	T_3 = .089
V_{4a} = 16363	V_{4b} = 18821	V_3 = 6710	T_4 = .495
		V_4 = 17560	w = −25.8′
Elmer:			
V_1 = 3000	V_{2b} = 5395	V_1 = 3000	T_2 = .650
V_{2a} = 5279	V_{3b} = 6882	V_2 = 5336	T_3 = .140
V_{3a} = 7023	V_{4b} = 19053	V_3 = 6948	T_4 = .636
V_{4a} = 16563		V_4 = 17697	w = 3° 00.2′
Lincoln:			
V_1 = 3000	V_{2b} = 6328	V_1 = 3000	T_{2a} = .039
V_{2a} = 6262	V_{3b} = 20438	V_2 = 6245	T_{3a} = .419
V_{3a} = 18798		V_3 = 19582	w = +2.26′
Norma:			
		V_1 = 3000	
V_{2a} = 5250	V_{2b} = 5228	V_2 = 5240	T_2 = .029
V_{3a} = 5912	V_{3b} = 5779	V_3 = 5830	T_3 = .049
V_{4a} = 7757	V_{4b} = 7755	V_4 = 7756	T_4 = .293
V_{5a} = 17494	V_{5b} = 18639	V_5 = 18040	T_5 = .704
			w = −43.8′
Millville:			
V_1 = 3000		V_1 = 3000	
V_{2a} = 5381	V_{2b} = 5452	V_2 = 5440	T_2 = .026
V_{3a} = 8044	V_{3b} = 7817	V_3 = 7926	T_3 = .353
V_{4a} = 16931	V_{4b} = 19514	V_4 = 18086	T_4 = .946
			w = 3° 21.02′
Port Elizabeth:			
		V_1 = 3000	T_1 = .035
		V_2 = 5344	T_2 = .353
		V_3 = 7637	T_3 = 1.088
		V_4 = 20233	
Woodbine:			
V_1 = 3000		V_1 = 3000	T_1 = .007
V_{2a} = 5445		V_2 = 5445	T_2 = .522
V_{3a} = 8613		V_3 = 8613	T_3 = 1.179
V_{4a} = 16639		V_4 = 16639	
Seaville:			
V_1 = 3000		V_1 = 3000	T_1 = .037
V_{2a} = 5589		V_2 = 5589	T_2 = .688
V_{3a} = 9850		V_3 = 9850	
Avalon:			
V_1 = 3000		V_1 = 3000	T_1 = .002
V_{2a} = 5495		V_2 = 5495	T_2 = .227
V_{3a} = 7073		V_3 = 7073	
Bridgeport:			
V_1 = 3000	V_1 = 3000	V_1 = 3000	
V_{2a} = 4450	V_{2b} = 4450	V_2 = 4450	T_2 = .008
V_{3a} = 17100	V_{3b} = 16000	V_3 = 17100	T_3 = .082

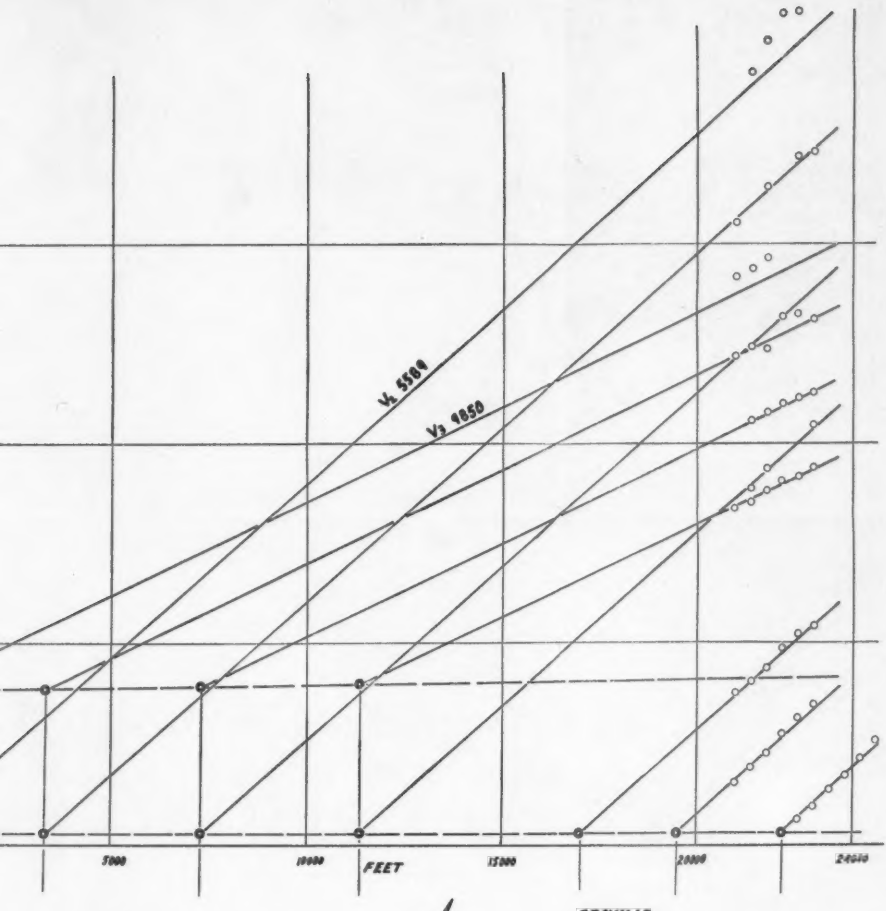
V = True velocities, ft./sec.
V_a and V_b = Reversed apparent velocities, ft./sec.
T values = Time intercepts, sec.
w values = Basement slope (horizontal angle)

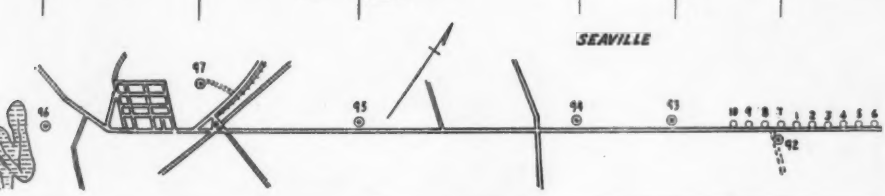


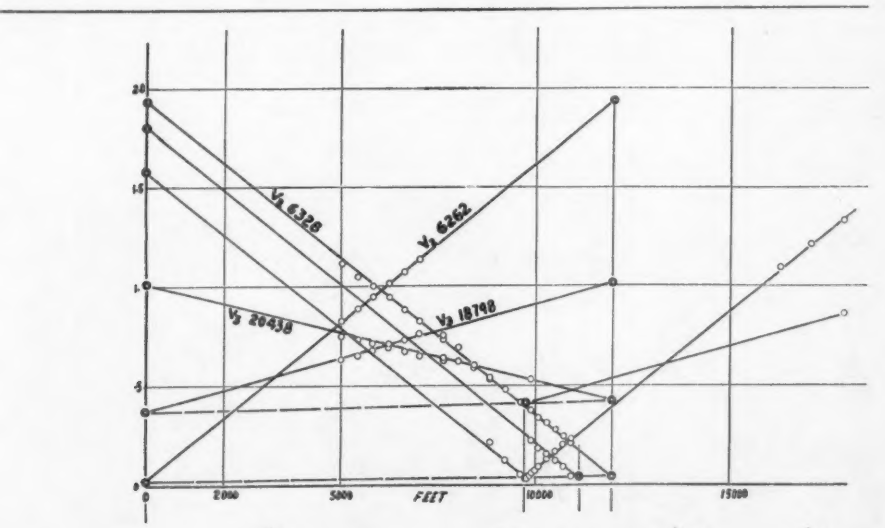


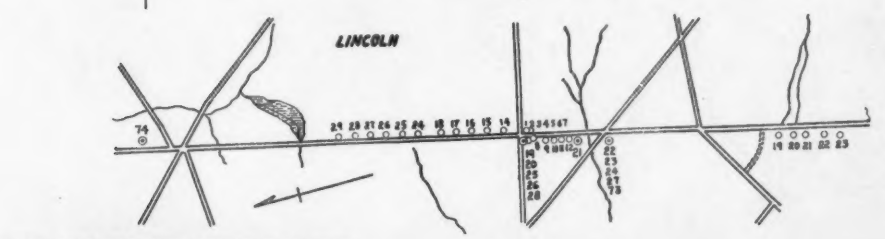


SEISMIC STATION PLOTS AND









LOTS AND TRAVEL-TIME GRAPHS

station. Despite charges up to 120 pounds of dynamite, energy failed to come through sufficiently from the basement to be recorded by the seismograph. This was strangely in contrast to the surface which transmitted the slightest disturbance over considerable distances. Part of the trouble was due to this later factor. In order to eliminate traffic noises from a main highway parallel to the station line at a distance of about three-quarters of a mile it was necessary to reduce the sensitivity. Despite the use of a charge of explosive about three times the normal amount for this distance, no waves from the basement were recorded. The only deep horizon defined here is the *"M"* horizon, for which there are abundant data.

AVALON

The Avalon station occupies practically the entire length of the offshore bar on which the town is located. The *"V"* horizon is well defined, but neither the *"M"* horizon nor the basement was detected. In the attempt to get a record from the basement, a charge of 83 pounds of dynamite was exploded about 5 miles from the geophone spread, but no ground motion of any sort was detected from this explosion.

BRIDGEPORT

This station was occupied twice, once with a long geophone spread and once with a short one. For the long spread the time-break is questionable, the data from it being plotted on the line marked 16000 on the travel-time graph. The parallel dashed line represents these same data after a correction of 0.032 sec. has been applied to make the time-intercept of Shot 78 agree with those from the other shots. The dashed line was used in the calculation of depth. It should be noted that the slope of the dashed line is concordant with the data from the other shots and that this slope was not affected by the correction applied to the time-break.

SUMMARY OF RESULTS

The velocities, slopes, and depths obtained at each station are summarized in Table 2. These depths are plotted on the geologic section in Plate 1.

Seismograms for Norma and Lincoln are reproduced in Plate 5.

GEOLOGIC SIGNIFICANCE OF RESULTS

GENERAL STATEMENT

As pointed out in the introduction, this line of seismic stations was established so as to yield the maximum amount of geological information in an area where there was little knowledge of subsurface structural conditions. Although the seismic results at Seaville and Avalon present

TABLE 2.—*True velocities and depths*

Station	"V_1"	"C"	"V"	"M"	"B"	H_1	H_2	H_3	H_4	H Total
Bridgeport	3,000	4,450			17,150	16	187			193
Prospect	3,000	5,510			17,598	3	283			286
Swedesboro	3,000	5,881			17,225	21	488			508
Lincoln	3,000		6,295		19,582	67	1,246			1,313
Pittsgrove	3,000	5,100	6,710		17,560	2	346	1,323		1,671
Elmer	3,000	5,336	6,948		17,697	91	354	1,698		2,143
Norma	3,000	5,240	5,830	7,756	18,040	53	275	939	1136	2,403
Millville	3,000	5,480		7,926	18,086	47	1,212	2,160		3,419
Port Elizabeth	3,000	5,344		7,637	20,233	63	1,171	2,573		3,807
Woodbine	3,000	5,445		8,613	16,639	13	1,806	2,773		4,592
Seaville	3,000	5,589		9,850		66	2,192			2,258
Avalon	3,000	5,495	7,073			4	973			977

V_1 = Surface gravel velocity, ft./sec.
"C" ..... V_2 = Cohansey formation velocity, ft./sec.
"V" ..... V_3 = Vincentown horizon velocity, ft./sec.
"M" ..... V_4 = Raritan-Magothy horizon velocity, ft./sec.
"B" ..... V_5 = Basement velocity, ft./sec.
H = thickness of respective velocity horizons, ft.
H total = depth to basement, ft.

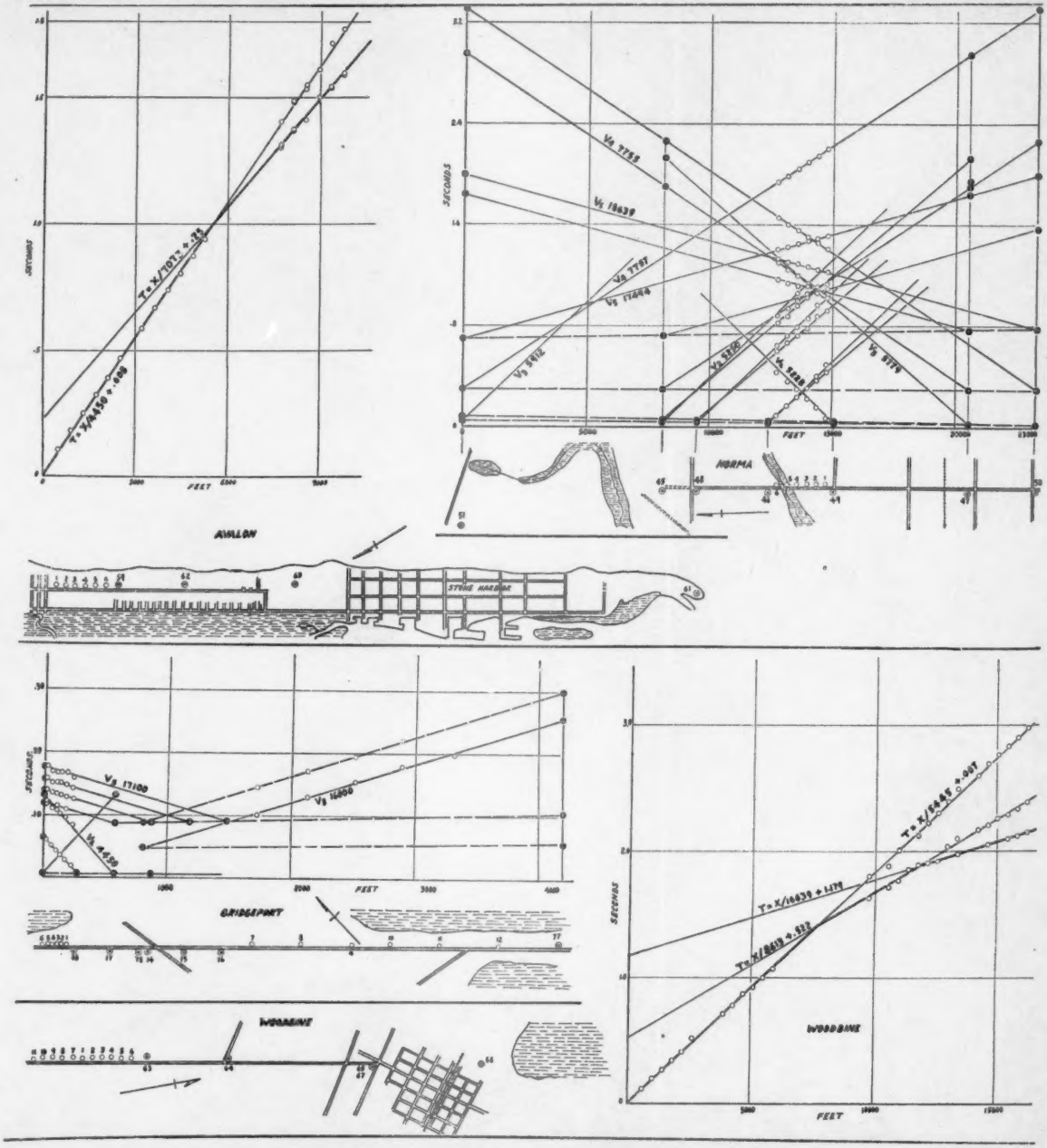
no evidence on the depth or nature of the basement, they do furnish information on nearer surface phenomena which is of interest.

GEOLOGY

The general geology of the New Jersey Coastal Plain and its underlying basement structure have been previously discussed in Part III (Ewing *et al.*, 1939, p. 276) and need no further comment here.

From the data supplied by 10 wells along the line of the traverse, and the geologic outcrops at the surface, it is possible to plot the position of the geologic formational equivalents of the *"V"* and *"M"* seismic horizons, on the shallow end of the section, and of the *"V"* horizon at a point about 22 miles from the coast where an abortive oil well was started near Cumberland, New Jersey. Most of the wells in the outer portion of the profile served only to establish the possible rate of dip of the upper *"V"* horizon by giving information on the shallower Kirkwood formation which is more or less conformable in dip with it.

As on the Barnegat Bay section, the *"V"* horizon agreed very closely



SEISMIC STATION PLOTS AND TRAVEL-TIME GRAPHS

in position throughout with the Vincentown formation. The *"M"* horizon, however, was detected only in the outer portion of the profile where there were no well data to permit a check. This correlation with the Raritan-Magothy unconformity here depends entirely on the similarity of the observed velocities with those for this discontinuity on the Barnegat Bay section, where well data permitted the identification, and on such other factors as observed velocities for the discontinuity, dip, and rate of thickening. All these lines of evidence confirm the identification.

TABLE 3.—*Velocity table*

(In feet per second)

Station	Surface gravels	Surface formation	*"V"* horizon	*"M"* horizon	*"B"* Basement
Bridgeport	3,000	4,450	.....	.....	17,150
Prospect	.....	5,510	.....	.....	17,598
Swedesboro	3,000	5,887	(5,887)*	.....	17,225
Lincoln	3,000	.....	6,295	.....	19,582
Pittsgrove	3,000	5,100	6,710	.....	17,560
Elmer	3,000	5,336	6,948	.....	17,697
Norma	3,000	5,240	5,830	7,756	18,040
Millville	3,000	5,480	.....	7,926	18,086
Port Elizabeth	3,000	5,344	.....	7,637	20,233
Woodbine	3,000	5,445	.....	8,613	16,639
Seaville	3,000	5,589	.....	9,850	.....
Avalon	3,000	5,495	7,073	.....	.....

* Surface formation here.

DISCUSSION OF OBSERVED VELOCITIES

The first column of velocities shown as 3000 feet throughout is the average value of the Cape May glacial gravels which cover almost the entire region. These vary in thickness from 1 or 2 feet to more than 50 feet. The "surface formation" refers to whatever geological formation underlies the gravels and would normally be exposed at the surface. At Swedesboro the surface formation is the *"V"* horizon associated with the Vincentown formation.

The *"V"* and *"M"* horizons have previously been discussed in connection with their geologic equivalents—the Vincentown formation and the Raritan-Magothy unconformity—and both horizons show a seaward increase in their respective velocity values. This agrees with the observations on both the Cape Henry, Virginia, and Barnegat Bay, New Jersey, profiles.

The variations in velocity observed for the "basement" do not show the same general systematic trend noted for the *"V"* and *"M"* horizons, and this is believed to be a reflection of the diverse changes in lithology encountered in crossing the basement.

COMPARISON OF CAPE MAY AND BARNEGAT BAY SECTIONS

In Plate 6 are shown comparative profiles for the two sections with the seismic horizons plotted relative to sea level along with their geologic extensions from well data. The Cape May section is characterized by a much greater rate of dip for the basement than the Barnegat Bay section. The comparative values taken along the base lines, which are dashed in to indicate the amount of relief in the basement surface, are 67 feet per mile on the Barnegat Bay section against 98 feet per mile on the Cape May section. Similarly, the amount of relief on the two sections varies from 285 feet maximum on the Barnegat Bay section to 490 feet maximum on the Cape May section.

In the overlying sediments the section lying below the *"M"* horizon in southern New Jersey is much thicker than in northern New Jersey, although the column lying above it is much the same in thickness. This increase in thickness in southern New Jersey probably indicates the presence of lower Cretaceous sediments which, although they do not outcrop in the State, are seen at the surface in Delaware. Another condition indicated on both profiles is the duplication of structure in the *"M"* and *"V"* horizons. Whereas the *"V"* horizon in both cases shows a relatively constant slope surface, the *"M"* horizon is characterized by a marked break and almost reversal of slope at Jackson's Mills on the Barnegat Bay profile and at Norma on the Cape May profile. As both of these localities lie over a ridge in the basement, this "break" in slope cannot be attributed to postdepositional folding, and the most probable explanation is that it represents an erosional feature in the Raritan-Magothy unconformity, the geologic counterpart of the *"M"* seismic horizon.

RELATIONS BETWEEN GRAVITY ANOMALIES AND SEISMIC VELOCITIES

General statement.—In 1937-1938 the gravity party of the U. S. Coast and Geodetic Survey established pendulum stations in the Coastal Plain area of New Jersey, Virginia, and North Carolina. Through the courtesy of the officials of the Survey, many of these stations were located along the seismic traverse lines to permit comparison of seismic and gravitational results. In Plate 6 the isostatic anomalies obtained from these gravity surveys are compared with the seismic velocities for the basement rocks.

As pointed out in Part III (Ewing, *et al.*, 1939, p. 283) a relation be-

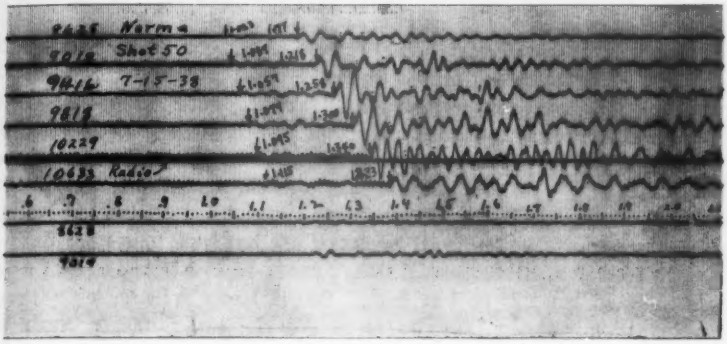


FIGURE 1. NORMA

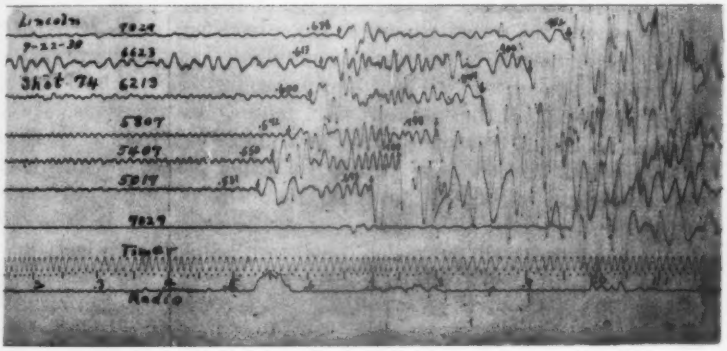


FIGURE 2. LINCOLN

TYPICAL SEISMOGRAMS FROM NORMA AND LINCOLN

Six detectors were used, the distance of each from the shot being indicated on the appropriate trace. The units of time indicated along the bottom of each seismogram are somewhat larger than seconds. The first arrival on each seismogram travelled through the basement, the second through the "V" horizon.

tween gravity anomalies and seismic velocities had been noted in Virginia. The same relation appears to hold in New Jersey. In general the areas characterized by low seismic velocities in the basement rocks are also characterized by minimum gravity anomalies and areas showing high seismic velocities show gravity maxima. Since the variations in seismic velocity and the pattern of the gravity anomalies both are controlled by changes in basement lithology, the writers have attempted to deduce the type of basement lithology from these two sets of data.

Both the Cape May and the Barnegat Bay profiles are characterized by a high velocity area of about 21,000 ft./sec. which is flanked on the west by velocities of about 18,000 ft./sec. and on the east by velocities of about 16,000 ft./sec. If the areas of equal velocity indicated on the two profiles are connected to form belts a strike is established which approximates that of the Appalachians, and studies of the gravity anomalies in the Coastal Plain have shown that gravity structures occur as belts parallel to the structure of the Appalachian Mountains (Woollard, 1938b, p. 1966). If the high velocity area is extended along the strike it is found to intersect the serpentine belt of Staten Island. However, H. H. Hess (personal communication) has called the writers' attention to unpublished studies on the position of serpentine belts in the eastern United States which indicate that the Staten Island serpentines do not pass under the Coastal Plain, but beneath the Triassic Basin to reappear in the Philadelphia area. The alignment of the high velocity areas with the Staten Island serpentines is so marked that the writers sought further evidence on the problem. It will be presented in the section on gravity measurements.

Information about the probable lithology of the other velocity areas comes from wells at Jackson's Mills in the 16,000 ft./sec. zone and Hightstown in the 18,000 ft./sec. zone. Rocks of these wells are believed to be variations of the Wissahickon schist, the Hightstown sample representing the higher degree of metamorphism. For the low velocity area east of the 16,000 ft./sec. zone of the Barnegat Bay profile no data were obtained on the Cape May profile, and for the 17,000 ft./sec. area of the Cape May profile there is no counterpart on the Barnegat Bay profile, for it presumably is intercepted by the Triassic Basin. Farther west, on the Cape Henry, Virginia, profile, the 13,000 ft./sec. area is believed to represent granite. This correlation, as pointed out in Part III (Ewing *et al.*, 1939, p. 285), does not agree with surface determinations of the velocity in granite, but the abundant well data in the area seem conclusive.

An areal section of the basement along the Barnegat Bay profile (Pl. 6) indicates the positions of the gravity stations in this area and the extent

of the lithologic units as revealed by seismic velocity measurements. There is not sufficient information to make the lithologic boundaries perfectly precise, but they are believed to be approximately correct.

Gravity anomalies.—Gravity anomalies represent the difference between an observed and a computed value of gravity. The computed

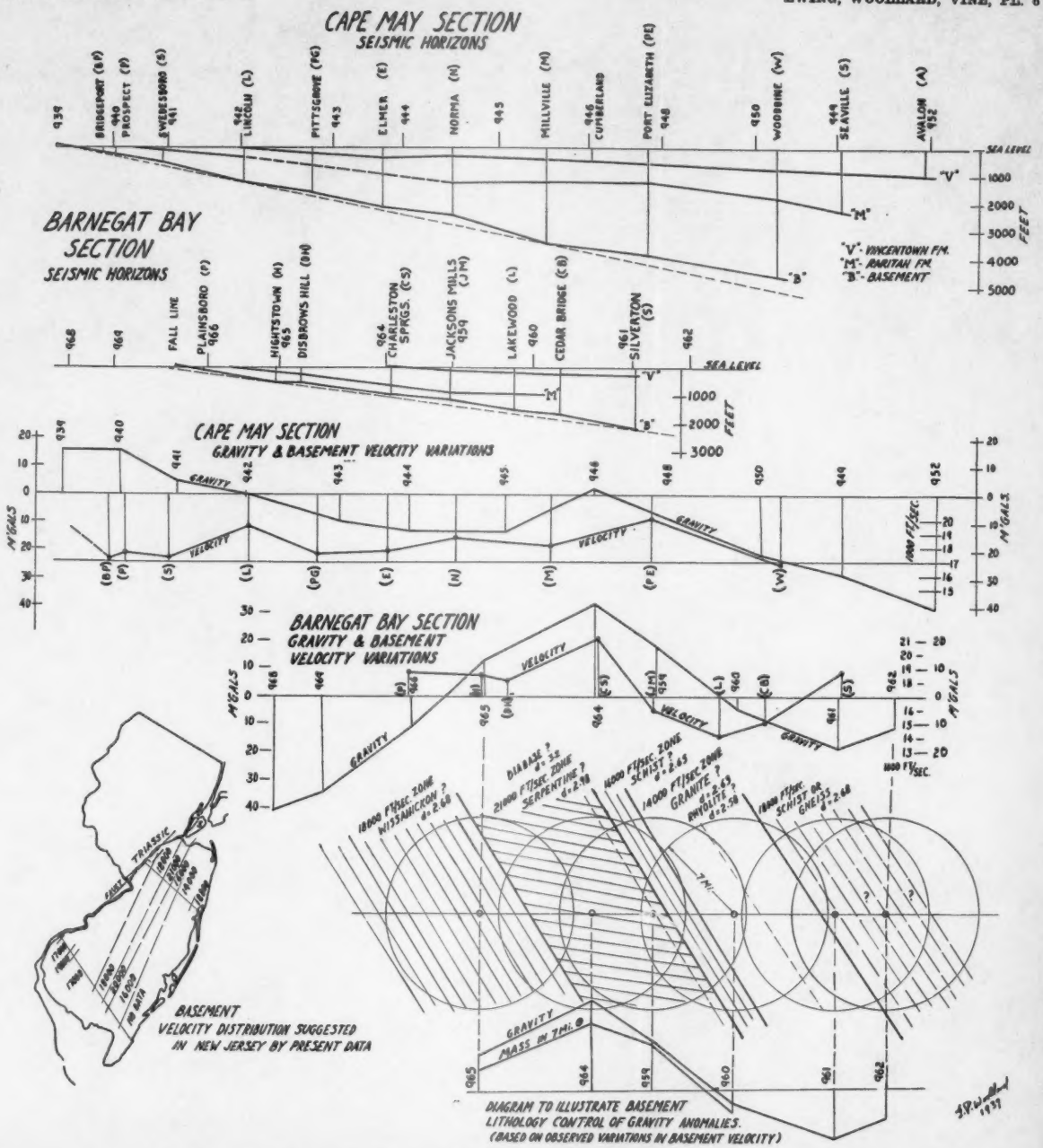
TABLE 4.—*Correlation between seismic velocities and rock types*

Velocity ft./sec.	Type of rock	Specific Gravity
18,000	Wissahickon gneiss	2.68
21,000	Serpentine or diabase or hornblendite	2.98 3.2
16,000	Wissahickon schist	2.65
14,000	Granite or rhyolite	2.63 2.58

value includes allowances for topography and for variations in density which may be considered as separated into parts due to conditions near the station and to regional effects. In the present traverse the effect of the overlying sediments may be considered to vary uniformly along the line, because the seismic measurements do not reveal abrupt changes in either the thickness or the composition of the sediments. The observed variations in gravity anomaly are considered due to variations in the density of the basement rocks, and calculations are made to find which of the possible types of rock would best explain the gravity anomalies.

In making a gravitational correction to be applied to the isostatic anomaly of each station, the tables of Cassinis (1937) were used for calculating the unit attraction out to Zone H (radius approximately 7 miles) and for a depth of 3.2 kilometers. The mean density of the rocks around each station was then calculated using the areal distribution of basement rocks shown in the section of Plate 6. The difference between this mean density and 2.67, when multiplied into the unit attraction, gave the correction. These corrections are similar to those previously applied to the anomalies in the Bighorn Region (Woollard, 1938a, p. 88).

On the basis of correlations between lithology and seismic velocities in Virginia, of the indications of seismic, gravitational, and geological observations in Virginia (Woollard, 1939, p. 320), and of unpublished magnetic observations along the Barnegat Bay line (Woollard, 1939), the correlations between lithology and seismic velocities shown in Table 4 are made.



COMPARATIVE DEPTH, SEISMIC VELOCITY, AND GRAVITATIONAL PROFILES

For the Barnegat Bay, New Jersey, and Cape May, New Jersey, traverses

In Table 4 there are two possibilities for the densities of the rocks with velocities of 21,000 ft./sec. and 14,000 ft./sec. To decide between the two possibilities in each case four calculations for the gravity correction were made using all possible combinations of rock types from Table 4.

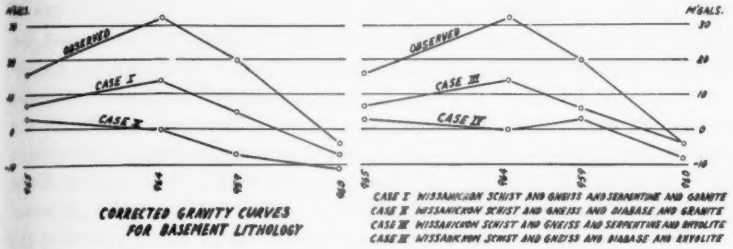

OBSERVED
CASE I
CASE II
CORRECTED GRAVITY CURVES FOR BASEMENT LITHOLOGY
OBSERVED
CASE III
CASE IV
MGALS.
CASE I WISSAHICKON SCHIST AND GNEISS AND SERPENTINE AND GRANITE
CASE II WISSAHICKON SCHIST AND GNEISS AND DIABASE AND GRANITE
CASE III WISSAHICKON SCHIST AND GNEISS AND SERPENTINE AND RHYOLITE
CASE IV WISSAHICKON SCHIST AND GNEISS AND DIABASE AND RHYOLITE

FIGURE 3.—*Residual gravity anomaly curves on Barnegat Bay traverse*

Corrected for basement lithology variations as determined from seismic variations. Case I Wissahickon schist and gneiss and serpentine and granite; Case II Wissahickon schist and gneiss and diabase and granite; Case III Wissahickon schist and gneiss and serpentine and rhyolite; Case IV Wissahickon schist and gneiss and diabase and rhyolite.

The combinations used together with the resulting corrected anomaly curves are shown (Fig. 3; Pl. 6). Of the four corrected anomaly curves in Figure 3, Case II is selected as best because it is free from the local anomalies which appear in the uncorrected curve. The uniform trend to negative anomalies in the eastward direction is considered to be a regional effect. This study of the gravity anomalies indicates that the velocities found in the seismic survey correspond to Wissahickon gneiss, diabase, Wissahickon schist, and granite.

The gravity anomalies at the west end of the Barnegat Bay profile are believed to be controlled by other factors than discussed above, because they lie in the Triassic area of the Piedmont Province and not in the Coastal Plain. Similar cases have been discussed for Virginia (Woollard, 1939, p. 322) and for New England (Longwell, 1938, p. 84).

There is not sufficient evidence for an explanation of the lack of correlation between seismic velocity and gravity anomaly at the most easterly station of the Barnegat Bay profile.

NEAR-SURFACE GEOLOGIC CONDITIONS ON CAPE MAY

As pointed out in the discussion of the various station profiles, it was impossible to obtain a record from the basement at the two outer stations—Seaville and Avalon—probably because they rest on a silted up lagoon. The area in question apparently lies in what must have been in Pleistocene time a lagoon behind an offshore bar similar to the one

now separating Barnegat Bay from the ocean. Apparently this silted up to form a peat area like that now lying between the present offshore bar at Avalon and the mainland.

Bennet (1937) shows that for the Cape May area, just south of the present section, the present lagoon lies on an erosion surface complete with a dendritic drainage pattern which is incised in the Cape May gravel surface of Wisconsin age; and he also indicates from well data that the repetition of gravels and fine clays suggests the presence of former lagoons separated by over-riding gravels. Similar indications are found in the log of the well at Cape May Court House and the log of the well at Rio Grande, where 40 feet of black peat muck lies at an elevation of −60 to −100 feet. The presence of clay rather than peat at Cape May Court House at a similar depth is believed to represent channel conditions as observed in the present lagoon where scour prevents the accumulation of peat.

The Cape May ridge, on which the Seaville seismic station is located, has no particular significance as to the position of a former offshore bar other than that it was the last location before some marked changes in sea level occurred in post-Wisconsin time. This is based on the history of the present offshore bar at Cape May, which since 1700 has migrated about three-quarters of a mile shoreward over its lagoon. This fact, coupled with the geologic evidence of successive stages of lagoons covered with gravels as seen in the Cape May section and the occurrence of peat outcropping on the seaward side of the present offshore bar, strongly suggests that the entire area may well be underlain by alternate deposits of gravels and peat. They would establish a natural absorber that would effectively prevent the vertical passage of energy, and this agrees with the seismic experience obtained in this area and appears to explain the failure to obtain results there.

CONCLUSIONS

In carrying out this project the writers used the standard method of seismic refraction technique and established reverse profiles wherever local cultural and natural conditions permitted. First and subsequent wave arrivals were used from the records in order to utilize all the information possible. To minimize interpretative errors, all of the data were least squared to arrive at the most accurate values for the time-distance graph velocities and intercepts. The method has been presented in the text.

The geologic indications of the investigation are best expressed in the following summary:

(1) The basement rocks dip seaward at an approximate rate of 98 feet per mile with an increasing gradient as the coast is approached.

(2) The velocity of the basement rocks varies between the same limits as observed by the writers in both Virginia and the central part of the New Jersey Coastal Plain, 14,000 ft./sec. and 21,000 ft./sec.

(3) The areal distribution of these velocity variations appears to be the same in the southern New Jersey Coastal Plain as in the central portion, suggesting that a banded lithologic control of seismic velocities extends between the two localities.

(4) Computations based on this hypothesis, whereby different rock groups were used possessing the seismic velocity values observed, showed that practically all of the gravity anomaly values observed over a test section of the seismic traverse could be removed, which further verifies the lithologic control of the observed seismic velocity variations.

(5) The rocks involved appear to be the Wissahickon gneiss, the Wissahickon schist, diabase, and porphyritic granite.

(6) The order of relief indicated by the seismic data for the basement surface is 400 ft. which is in agreement with that observed on previous investigations across the Coastal Plain.

(7) Seismic horizons in the overlying sediments, correlated from well and outcrop data with the Raritan-Magothy, Vincentown, and Cohansey formations, can be traced with a marked degree of accuracy across the Coastal Plain. These findings agree with those observed on the previous traverses conducted by the writers, although the Cohansey horizon had not been detected before.

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ARCHEAN METACONCRETIONS OF THUNDER LAKE, ONTARIO

BY F. J. PETTIJOHN

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ABSTRACT

Silicated calcareous concretions occur in paragneisses and mica schists of Archean age near Thunder Lake in northwestern Ontario. Anorthitic plagioclase, quartz, and hornblende are the chief constituents, and clinozoisite and garnet are the most significant minor components. The high lime content shown by chemical analysis, the peculiar mineral assemblage, the nodular structure of these bodies, and their relation to the containing beds leave no doubt of their concretionary origin.

Similar metaconcretions have rarely been reported from Archean rocks. Their importance in support of the uniformitarian principle is discussed.

INTRODUCTION AND ACKNOWLEDGMENTS

The importance of recording those features of the Archean rocks which are clues to the conditions during the earliest times is self-evident. The writer noted with interest, therefore, in 1937, nodules in the paragneisses and quartz-biotite schists of the Thunder Lake area of northwestern Ontario. These were interpreted as metamorphosed calcareous concretions (Pettijohn, 1939). A search of the literature on the early pre-Cambrian of the Canadian Shield failed to disclose any reference to similar metamorphosed concretions. Accordingly, a further description of these nodules and a discussion of their significance seemed desirable. The gen-

eral absence of limy sediments in the early pre-Cambrian—except the Grenville if, indeed, that be early pre-Cambrian—adds significance to the silicated calcareous nodules.

The expense of collecting materials for study, of chemical analyses, and of thin sections was defrayed by a grant from the Penrose Bequest of the Geological Society of America. This aid is gratefully acknowledged. Mr. Arthur Lundahl rendered able field assistance. Professor N. L. Bowen kindly read the manuscript and made helpful suggestions.

REVIEW OF LITERATURE

The occurrence of similar nodules in rocks of other places and ages has been noted by a few geologists. Keith (1913) early noted such nodules in both Cambrian and pre-Cambrian rocks in the southern Appalachians. To these Keith gave the name "pseudo-diorite" rather than diorite which they resembled in mineral composition. Keith correctly attributed them to "mineral recrystallization by solutions under pressure" rather than to true igneous activity. He noted the occurrences of these ellipsoidal bodies in metapsammitic beds and briefly described their appearance, distribution, mineral composition, and concentric structure. He failed, however, to relate them to calcareous concretions.

Laney in 1926, however, more fully described the same "pseudo-diorite" nodules earlier noted by Keith. More complete petrographic and chemical study led Laney to the conclusion that they were metamorphosed calcareous concretions.

Sederholm (1928) describes nodules from the pre-Cambrian mica schists of the Tulola Island in Lake Ladoga in Finland. Though they tally in all important respects with those described by Laney, Sederholm believed that "there is little doubt that they are due to the secondary action of the granitic magma" and that they were in some way related to the nodular (orbicular) granites.

Eskola (1932) gave a more ample description of these and other nodules from the Karelian schists of the Finnish pre-Cambrian and interpreted the nodules as metamorphosed calcareous concretions. According to Eskola this interpretation was accepted by Sederholm and other workers. Eskola noted the occurrence of these nodules in varved pre-Cambrian schists. They occur only in the thicker more silty or sandy beds. Mineralogically they are highly variable and can, according to Eskola, be arranged in a series representing increasing temperature and higher metamorphic grade. This series is:

(1) Calcite-quartz-epidote
(2) Calcite-quartz-epidote-grossularite
(3) Calcite-quartz-grossularite-actinolite

(4) Quartz-grossularite-plagioclase-actinolite-calcite
(5) Quartz-plagioclase-actinolite-grossularite
(6) Quartz-plagioclase-hornblende-grossularite
(7) Quartz-plagioclase-hornblende-diopside

The plagioclase varies in composition but tends to be high in lime, often bytownite or bytownite-anorthite. The lime- and alumina-rich silicates, such as anorthite, grossularite, and epidote, are interpreted by Eskola as reaction products between primary clayey materials and calcium carbonate.

More recently Runner and Hamilton (1934) have given a detailed description of nodules from some pre-Cambrian schists of the Black Hills, South Dakota. These nodules are triaxial ellipsoidal masses of lime plagioclase-grossularite-diopside-quartz composition. Their matrix is a siltstone now metamorphosed to quartz-feldspar-biotite schist. These bodies are almost identical in their field relations, structure, and mineral composition with those of Thunder Lake described in this paper. They were interpreted by Runner and Hamilton as metamorphosed calcareous concretions.

Most recent discussion of pre-Cambrian concretions is that of Campbell and Maxson (1935), who describe metaconcretions from the Vishnu (Archean) schists of the Grand Canyon of Arizona. These bodies, similar in mineral composition to those described by other writers, still contain some of the original carbonate.

FIELD OCCURRENCE

The concretions herein described occur in quartz-biotite-feldspar schists or paragneisses of Archean age near Thunder Lake in northwestern Ontario (Pettijohn, 1939). They are especially abundant in outcrops along the north line of Zealand township, lots 11 and 12. The country rock was originally a series of alternating siltstones, graywackes, and slates. Though original structures and textures are rather well retained, metamorphism has converted the series to schists differing from one another in the proportions of quartz, feldspar, and biotite as well as in grain size.

The nodules occur only in the thicker, originally sandy beds. They vary from small nearly spherical bodies about 2 inches in diameter to lenticular, layerlike units as much as 36 inches long. Such larger bodies are irregular and appear to have formed by the union of several smaller nodules. A few nodules have a dumb-bell form, obviously the result of coalescence of nodules formed independently, but most commonly a simple flattened ellipsoidal form is seen (Pl. 1, fig. 1).

The nodules resist weathering better than the enclosing matrix and are, therefore, etched into relief by weathering of the outcrops. A distinct

zoning was noted in some of the nodules. Commonly but two zones can be recognized; less commonly three or even four can be seen. Except for this rude zoning no internal structure, such as schistosity, was noted. In some few instances, however, the bedding of the schist was observed

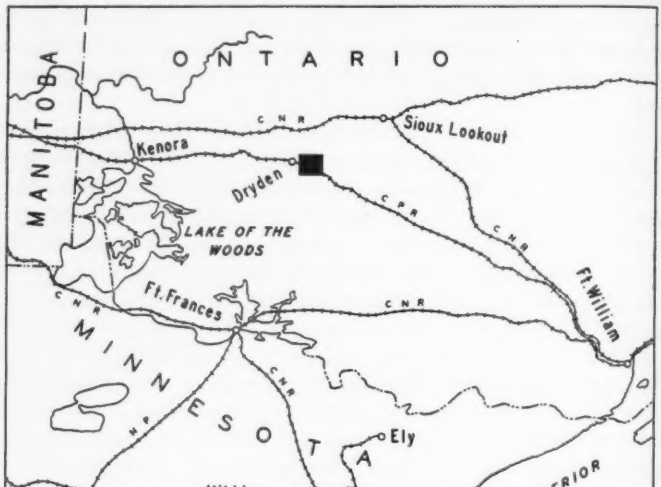

FIG. 1.—*Map showing location of Thunder Lake area*

to pass through the concretions, indicative of an epigenetic origin of the latter.

PETROGRAPHY OF THE METACONCRETIONS AND MATRIX

The strata in which the nodules are found are well bedded. They have been recrystallized but are in general without conspicuous cleavage. When freshly broken, the matrix is a fine-grained gray rock; when weathered, it assumes a buff color.

The nodules themselves are more coarsely crystalline than the matrix, and the abundance of large unoriented hornblende crystals gives the material a dioritelike appearance.

Under the microscope the nodules show a granoblastic matrix of quartz and plagioclase with large poikiloblastic crystals of hornblende, together

with minor amounts of garnet, clinozoisite, titanite, chlorite, carbonate, and pyrite.

The quartz, in small rounded to subangular grains .05 to .2 mm. in diameter, forms about 40 per cent of the nodules. The quartz is free from strain effects and is generally clear and free from inclusions. The grains are for the most part the original quartz grains of the sediment in which the nodule was formed.

The feldspar, in smaller grains on the average than quartz, forms about 30 per cent of the nodules. It occurs between and molded around the latter and in a few cases completely envelops the quartz. A rude kind of net work of feldspar is thus formed in which the "holes" in the net are the quartz grains. The feldspar is plagioclase marked in most cases by a lack of twinning and by its conspicuous relief. Its indices are somewhat greater than those of quartz. Determination with immersion liquids yields refractive indices which correspond to those of bytownite-anorthite (An_{85}). Calculations based on the chemical analysis of the nodule lead to a similar conclusion.

The hornblende, in large poikiloblastic grains about 4mm. in length, forms 20 to 25 per cent of the metaconcretion (Pl. 2, fig. 1). The grains are elongate, with random arrangement, and marked by frayed ends. Inclusions are numerous and in some cases constitute as much as one third of the entire grain. Most common are quartz and plagioclase, but some clinozoisite, pyrite, carbonate, and chlorite are also present in the hornblende. The hornblende is clearly later than the quartz and feldspar and replaces these constituents. It is pleochroic X = greenish yellow, Y = yellow green, Z = blue green. To a slight extent the hornblende has been altered to chlorite.

Garnet occurs in sievelike patches visible to the naked eye, 1 or 2 mm. across, and forms less than one per cent of the total (Pl. 2, fig. 2). Inclusions are numerous and are mainly quartz. The peripheral parts of these cellular grains show straight euhedral contacts with the quartz-plagioclase mosaic of the nodule matrix. The garnet is pale pink and is probably a calcium-aluminum garnet with a little iron, nearest grossularite in composition. It is clearly later than the quartz and feldspar.

Clinozoisite occurs in numerous small euhedral, obliquely terminated, lath-shaped prisms, less than 0.1 mm. long, scattered singly or in clusters of two or three throughout the nodules. The clinozoisite prisms show an anomalous "berlin-blue" color, though some are zoned and show a first order white or even yellow in the iron-rich periphery. The clinozoisite cuts across and apparently replaces both quartz and feldspar, most especially the latter since it is generally molded around the quartz but tends to cut through the feldspar (Pl. 2, fig. 3). Clinozoisite does not

appear in either the hornblende or garnet except in the parts of the hornblende replaced by chlorite.

Sphene occurs as scattered rounded to irregular granules of small size replacing the feldspar. Other minor constituents are chlorite, which is closely associated with the hornblende and apparently formed by alteration of that mineral, a few scattered grains of pyrite, and a little carbonate which is probably secondary and not an unaltered residue from the original concretion.

The matrix of the nodules is what the British petrographers call a "biotite granulite." It is neither coarse enough nor banded well enough to be called a gneiss nor fissile enough to be classified as a schist. Yet it is rich in feldspar—more feldspar than quartz—and the micas are well enough oriented to give the rock a flaggy parting parallel to the original bedding.

Under the microscope it is seen to consist of a granoblastic matrix of quartz and an untwinned plagioclase in which are scattered oriented flakes of biotite and, to a lesser extent, muscovite (Pl. 2, fig. 4). The feldspar is andesine. The biotite is the brown variety, strongly pleochroic from straw yellow to dark brown. Muscovite and chlorite are closely associated and intergrown with the biotite, and the latter is probably derived from these two minerals. A little secondary carbonate, pyrite, sphene, and apatite form the minor accessory minerals. No orthoclase was observed. Very probably the potash is wholly in the micas.

The original sediment was undoubtedly a fine-grained sandstone rich in feldspar and containing enough impurities to form biotite upon metamorphism. The lime which formed the concretions was segregated in the nodules prior to metamorphism.

In the same outcrops in which the nodules are found, one may observe layers but a few inches thick of material superficially similar to the nodules in mineral composition. The layers likewise resist weathering better than the associated paragneiss and are therefore etched into relief (Pl. 1, fig. 2).

The apparent similarity in composition is confirmed by microscopic examination. The layers, like the nodules, consist of a dominant granoblastic "matrix" of quartz and anorthitic plagioclase in which are randomly oriented poikiloblastic crystals of a blue-green hornblende. The feldspar is less basic than that of the nodules but it is nevertheless near labradorite in composition. Some specimens contain a little garnet, others contain clinozoisite or a little biotite. The rock may properly be called a hornblende granulite or a biotite-hornblende granulite. The close resemblance of this rock to the nodules in both texture and mineral

FIGURE 1. DISCOID CONCRETION IN BIOTITE GRANULITE MATRIX
Note two zones.

FIGURE 2. OUTCROP OF ARCHEAN (ABRAM) GRANULITE
Showing silicated calcareous layers. The latter are more resistant and stand up in relief.

FIELD OCCURRENCE OF METACONCRETIONS

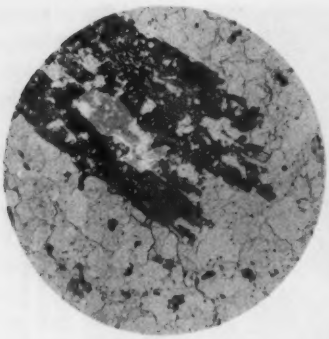

FIGURE 1. POIKILOBLASTIC AND "FEATHERED" HORNBLENDE
Note clear quartz surrounded by dusty bytownite. Dark areas in hornblende are haloes around minute inclusions. x 29.

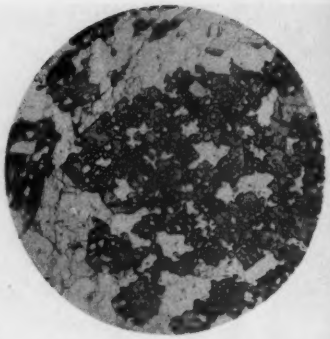

FIGURE 2. SIEVELIKE GARNET METACRYST
Note its tendency to assume straight, sharp boundary against quartz-bytownite matrix. Inclusions mainly quartz. x 29.

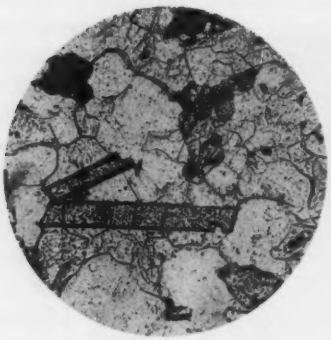

FIGURE 3. CLINOZOISITE PRISMS REPLACING BYTOWNITE
Small "embayment" near end of larger clinozoisite is result of growth around small quartz grain. Both crystals lie in bytownite and at each end against quartz. Note rounded quartz (original detrital grains) and cleavage of interstitial feldspar molded against quartz. x 175.

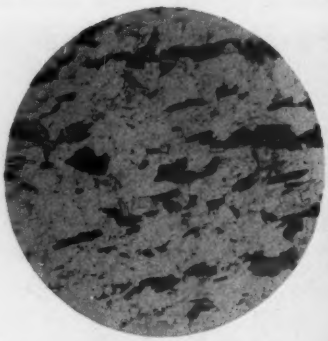

FIGURE 4. QUARTZ-BIOTITE SCHIST MATRIX
Mainly oriented brown biotite (with some muscovite and chlorite) in granular quartz-feldspar "matrix". x 29.

PHOTOMICROGRAPHS OF METACONCRETIONS
All under ordinary light

composition leads to the conclusion that, like the nodules, it has been formed by the recrystallization of a calcareous arenite or grit.

CHEMICAL COMPOSITION

Samples were taken of both the nodules and their matrix for chemical analyses. For comparative purposes some published analyses of similar materials are included in Table 1.

TABLE 1.—*Chemical Analyses*

	A	B	C	D	E	F	G
SiO_2	68.13	66.65	66.20	67.11	51.10	58.30	23.92
Al_2O_3	13.88	16.78	15.08	15.44	10.20	22.99	6.82
Fe_2O_3	0.77	0.36	0.04	0.89	0.12	0.74	1.12
FeO	3.83	2.93	2.94	3.89	2.85	1.00	2.57
MgO	1.68	1.86	1.12	0.19	0.54	1.51	2.26
CaO	9.04	3.25	12.31	6.79	17.92	14.00	31.35
Na_2O	0.62	4.29	0.77	2.58	0.53	n.d.	1.16
K_2O	0.08	1.69	0.07	0.41	2.00	n.d.	1.84
H_2O+	0.50	0.87	0.49	0.70	1.10	0.29	2.04
H_2O-	0.04	0.04	0.08	0.03	0.50	0.08	0.64
CO_2	0.27	0.49	0.53	.....	12.20	.....	24.24
TiO_2	0.70	0.39	0.47	0.34	0.38	0.50	0.62
P_2O_5	0.16	0.11	0.23	0.26	T	.....	0.36
MnO	0.19	0.07	0.18	0.43	0.42	0.20	0.93
SO_3	.....	.....	.....	0.13	.....	.....	.....
	99.89	99.78	100.51	99.19	99.86	99.61	99.87

A. Metaconcretion from Abram series, Thunder Lake, Ontario, R. B. Ellestad, analyst.
B. Paragneiss matrix of "A." R. B. Ellestad, analyst.
C. Concretion in mica schist, Tulolansaari, Sortavala, Finland. (After Hackman, 1931.)
D. Pseudodiorite, mine No. 20, Fannin County, Georgia. Chase Palmer, analyst. (After Emmons and Laney, 1926.)
E. Metamorphosed calcareous concretions occurring in the quartz-sericite schists of Boucher Creek (Vishnu). W. H. Herdsman, analyst. (After Campbell and Maxson, 1935.)
F. Analysis of the inside of an ellipsoid from pre-Cambrian sedimentary schists, Black Hills, South Dakota. H. Benninghoff, analyst. (After Runner and Hamilton, 1934.)
G. "Imatra stone," calcareous concretions in late Glacial clay, Imatra, East Finland. L. Lokka, analyst. (After Eskola, 1932.)

Inspection of the analysis of the Thunder Lake nodules (A) shows a high lime, alumina, and silica content, as would be expected from the mineralogical composition. As noted by Bowen, anorthitic plagioclase and abundant quartz form a combination which does not occur in igneous rocks, so that a magmatic origin for these materials is improbable.

Comparison of the analysis of the concretion and that of its matrix

shows the nature of the segregation process involved during the formation of the concretion. The rarer minor constituents, notably MnO and P_2O_5, are higher in the nodules than in the matrix. This contrast is also true of the pseudodiorite and enclosing graywacke described by Emmons

TABLE 2.—*Calculated mineralogical composition*

	Nodule (A)	Matrix (B)
Quartz	42.3	31.3
Albite	5.2 } An_{85}	36.2 } An_{23}
Anorthite	30.0 }	10.8 }
Hornblende*	18.6	....
Titanite	1.8	1.0
Calcite	0.6	1.1
Apatite	0.5	0.3
Biotite†	....	19.1
Total	99.0	99.8

* Calculated as actinolite to which excess alumina (1.74%) was added.
† Calculated as biotite but includes some chlorite and muscovite.

and Laney and of the metaconcretion and its mica schist matrix described by Eskola. These constituents are also high in the Quaternary concretions (G). It is apparent from the high lime content, however, that the concretions were essentially lime concretions and that the lime- and alumina-rich silicates, in particular the bytownite-anorthite feldspar, are the reaction products between the primary clayey minerals and the calcium carbonate. The CO_2 has been expelled from the Thunder Lake nodules.

The analysis of the Thunder Lake metaconcretion is singularly similar to that of the concretion (C) from the Archean mica schist of Finland. The Thunder Lake material is also similar to the pseudodiorite (D) and to the metaconcretion from the Archean of the Black Hills (F). Such differences as exist may have been determined by difference in the composition of the parent rock. The Vishnu concretions (E) differ from all the others in that they still retain considerable carbonate.

The matrix (B) has the composition of a typical graywacke. It has one third as much lime as the concretions but has, on the other hand, a notably greater alkali content. The latter suggests a high feldspar content in the original sediment.

The measured mode has essentially the same mineral frequencies as those calculated from the chemical composition. The most important

exception in the nodule, however, is the presence of 7.7 per cent of clinozoisite not disclosed by calculation. The clinozoisite was apparently formed at the expense of the feldspar since the observed amount of feldspar is but 27.3 per cent, whereas the calculated amount is 35.2.

ORIGIN AND SIGNIFICANCE OF METACONCRETIONS

The nodules, the published descriptions of which have been cited, are very similar to one another and in no important detail different from the material of Thunder Lake. The origin of these bodies, once not so obvious, now seems well established. The nodular form, the zoned structure, the occurrence in metapsammitic beds, the association of basic bytownite and other lime silicates with quartz, and the high lime content shown by chemical analyses all support the idea that these dioritelike bodies are the product of metamorphism of calcareous concretions. An epigenetic origin is indicated by the passage of bedding planes through some of the nodules.

The occurrence of the nodules in the coarser beds only is perhaps due to the greater permeability of these beds to ground-water flow. A similar relationship exists between the calcareous nodules and the thicker sandy varves of the lake sediments associated with the retreat of the late Wisconsin ice.

The general deficiency of limy sediments of the Archean has been pointed out. The concretions, however, show that lime was as freely transported and deposited by ground waters as now. The hornblende granulite beds associated with the nodule-bearing schists were primary calcareous grits and show conclusively that conditions for lime deposition in the Archean did exist. The lack of limy sediments, therefore, cannot be explained in an "exceptionalistic" manner. The writer (1940) has elsewhere suggested that the scarcity of limestone and also quartzite might be explained by the "mutual incompatibility" of these types and the geosynclinal or flysch facies. The former, the products of sedimentation in epicontinental seas, are not present in the axial part of the geosyncline and hence not preserved in the deepest downfolded part of that belt.

The habitual association of true concretions and organic remains in later sediments is well known. No direct evidence of such association, however, is forthcoming from the Archean counterpart unless the graphitic nature of some of the metapelites be so considered.

In conclusion, the Archean metaconcretions of Thunder Lake, like those described by Eskola in Finland, present evidence of the validity of the principle of uniformitarianism.

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GLACIAL CHRONOLOGY OF THE SOUTHERN ROCKY MOUNTAINS

BY LOUIS L. RAY

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ABSTRACT

In order to extend the chronology and validate the five substages of the Wisconsin glaciation recognized in the Cache la Poudre Valley in the Colorado Front Range, a reconnaissance was made of the Southern Rocky Mountains, from southern Wyoming to Santa Fe, New Mexico. The chronology determined in the Cache la Poudre Valley was used to date the culture layer of the Lindenmeier (Folsom) Site in northern Colorado.

The earliest or Twin Lakes substage is named for the "early moraine" mapped by Capps at Twin Lakes in the Upper Arkansas Valley. This substage is believed to have been contemporaneous with the Durango glaciation of the San Juan Mountains.

Throughout the Southern Rocky Mountains the second, third, and fourth, or Home, Corral Creek, and Long Draw substages, can be readily correlated from valley to valley. The fifth or youngest substage, represented in some cirques by protalus ramparts, is named the Sprague substage.

The validity of five distinct substages of Wisconsin ice advance, separated from one another by interstadial ice retreat or complete disappearance, is based on: (1) the character and topographic position of the moraines; (2) the relationship

between successive terraces (valley trains) and the terminal moraines, as determined in the Cache la Poudre Valley; (3) the differences in weathering of the tills composing the moraines; and (4) the reported finding of an old soil zone between tills assigned to the Twin Lakes and Home substages.

INTRODUCTION

GENERAL NATURE OF PROBLEM

Numerous archeological discoveries in Europe, Asia, and Africa, as well as in North America, have in recent years opened an important field for co-operation between the archeologist and the geologist. Culture layers of archeological sites were deposited and preserved as a result of geological events which must be fitted into the geological chronology. To the geologist falls the task of determining that chronology and interpreting the physical history of archeological sites.

In North America no co-ordination has been attained whereby, as in Europe, late geologic events can be interpreted on the basis of archeological chronology. In western United States, numerous archeological sites await geologic interpretation. Past erosional, depositional, and climatic history must be regionally studied and rationally interpreted. Into the geological history must be fitted the chronological sequences of fossil plants and animals as well as the artifacts of ancient man.

With an opportunity for co-operation and correlation between geology and archeology, the writer, in 1936 and 1937, worked with the Smithsonian Institution at the Lindenmeier Site in northern Colorado. Here, artifacts of the Folsom Culture are imbedded in a definite culture layer. The purpose of the field work was to determine the age of that layer. At this site the artifact-bearing stratum is associated with a terrace which is correlated with a glacial moraine in the near-by Cache la Poudre Valley of the northern Colorado Front Range (Bryan and Ray, 1940).

During the study of the Cache la Poudre Canyon five subdivisions of the Wisconsin stage of glaciation were recognized, instead of a single Wisconsin ice advance, previously recognized throughout the Southern Rocky Mountains. Distinct and separate moraines established the existence of three Wisconsin substages; an inferred moraine indicated an earlier substage, and a protalus rampart within the cirque indicated a late period of climatic refrigeration when great snowbanks lay in the cirques but no glaciers were developed.

To prove that these successive Wisconsin glaciations are not unique to the drainage basin of the Cache la Poudre River, a reconnaissance of selected portions of the glaciated mountains of Colorado and New Mexico was made in the summer of 1938. The division of the Wisconsin stage into substages will have a far-reaching effect on the regional correlation and on the dating of many Pleistocene to recent events. High stands of

lakes in the Basin and Range province, the dates of river terraces produced by climatic changes, and many other groups of phenomena that can be correlated with the fluctuating climate of the Pleistocene are referable to this newly established Wisconsin subdivision. Pollen analysis and various ecological and paleontological studies must reflect the complexity of the chronology and must be fitted into this more complicated history of the Wisconsin glaciation.

CHARACTER OF FIELD WORK

An examination of representative glaciated areas in the Southern Rocky Mountains was made to determine the validity of the subdivision of the Wisconsin stage of glaciation. The areas examined lie between the Medicine Bow Mountains of southern Wyoming and the southern limit of glaciation in the Sangre de Cristo Mountains, near Santa Fe, New Mexico (Fig. 1).

Of necessity, areas of study were chosen for two reasons. (1) They were critical because of location or because of previous studies; (2) they were easy of access. Figure 1 shows their general location.

During the field season of 1938 about 30 cirques were entered, numerous valleys studied, and a trip made to the San Juan Mountains where the well-known Durango and Cerro tills (Atwood and Mather, 1932) were examined. In some valleys it was feasible to spend only a day; in others several days. Inclement weather in the higher mountains during the summer and heavy rains and floods of early September impeded the work. However, enough material was gathered to show that the Wisconsin stage in the Southern Rocky Mountains consists of several definite periods of ice advance and recession. Without doubt, further study will reveal still greater complexity in the late glacial history, but the writer believes it unlikely that more than five widespread substages of the Wisconsin stage will be established.

ACKNOWLEDGMENTS

This reconnaissance was financed by a grant from the Milton Fund of Harvard University. Professor Kirk Bryan sponsored the project and has given many helpful suggestions, both in the office and in field conferences. For careful criticism of the manuscript the writer is especially grateful to Professor Bryan and Mr. A. P. Butler of Harvard University.

The efficient and volunteer field assistance of Mr. W. M. Johnson of Amherst College is gratefully remembered. The aid and company of Dr. H. T. U. Smith of the University of Kansas, Dr. J. Fred Smith, Jr. of Texas Agricultural and Mechanical College, Dr. Charles S. Denny of Dartmouth College, and Mr. E. C. Cabot of Harvard University on vari-

ous climbs into the high cirques was unique and inspiring. Various officers of the United States Forest Service and National Park Service aided the writer through their intimate knowledge of the country. The officers of

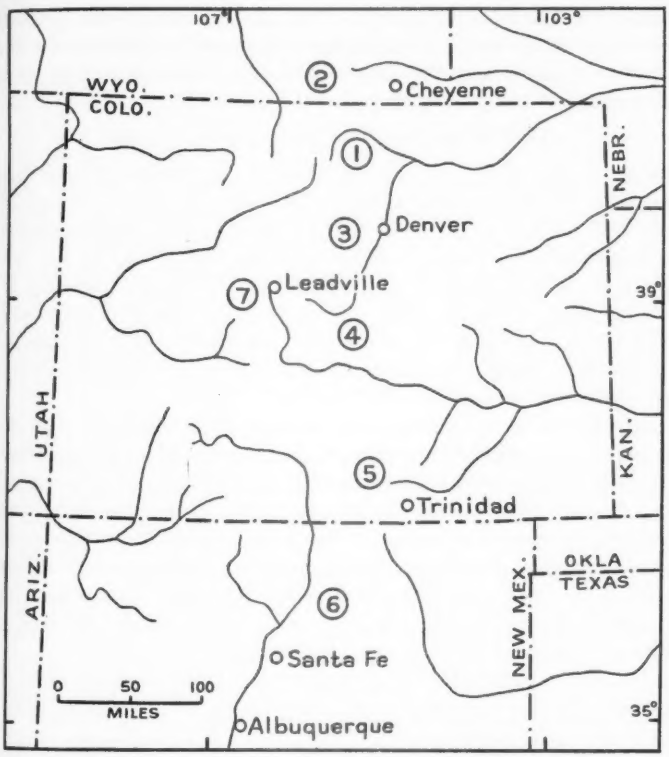

FIGURE 1.—*Index map showing location of glaciated regions studied*

(1) Cache la Poudre drainage basin and Rocky Mountain National Park; (2) Medicine Bow Mountains (Snowy Range); (3) Mount Evans and adjacent regions; (4) Pikes Peak region; (5) Blanca Peak, Spanish Peaks, and Culebra Range; (6) Sangre de Cristo Mountains of New Mexico; (7) Leadville and Aspen region.

Rocky Mountain National Park were particularly helpful. When in the Medicine Bow Mountains, the writer was a guest at the University of Wyoming Summer Camp. Many local residents of both Colorado and New Mexico perpetuated the tradition of the west with freely tendered aid and hospitality.

CACHE LA POUDRE DRAINAGE BASIN

GENERAL STATEMENT

A preliminary account containing part of the glacial chronology developed during the study of the glacial history of the Cache la Poudre drainage basin was briefly reported at the Philadelphia Symposium on Early Man (Bryan, 1937) and at the symposium on the physiographic development of the Front Range (Ray, 1938). Recently, the results of this study have appeared in more detail (Bryan and Ray, 1940).

PRAIRIE DIVIDE STAGE (PRE-WISCONSIN) GLACIATION

The oldest glacial deposit found in the Cache la Poudre drainage basin is at Prairie Divide, a broad mountain flat with an altitude of approximately 7900 feet (T. 10 N., R. 72 W., Livermore and Home quadrangles). Here, weathered gravel rests on deeply weathered till (Pl. 1, fig. 1). They form a plain so perched above the level of the present drainage that it is now being eroded from all sides. Large bodies of slumped material fill valleys which have cut headward into the mass. Small patches of weathered loess containing calcareous concretions lie within the slumped masses, below the general level of the plain.

Prairie Divide is a remnant of a widespread glaciation. The deep weathering and topographic position of the till above the valleys containing the more recent glacial deposits testify to the antiquity of this glacial stage. Prairie Divide is about 8 miles east of the most advanced position of the later Wisconsin moraines, a fact which points to a much greater extension of the glaciers during this stage than in later times. This early glaciation has been named the Prairie Divide stage (Bryan and Ray, 1940) and is thought to be correlative with the Cerro glaciation of the San Juan Mountains (Atwood and Mather, 1932).

Following the Prairie Divide glaciation there was a long period of erosion—the "Canyon Cutting Cycle" (Van Tuyl and Lovering, 1935)—when most of the glacial deposits were removed from the surface over which the Prairie Divide glaciers had moved. This period of valley deepening indicates long continued erosion.

The refrigeration of climate attended by the advance of the Wisconsin glaciers into the valleys cut during the "Canyon Cutting Cycle" developed a severe periglacial climate whose effects are evident at Prairie Divide. Valleys cut below the glacial debris are choked with slumped till and gravel which radiate outward from the central mass as long tongues of debris.

Striking evidence of the periglacial climate is found in the nivation cirque developed in the till and gravel on the eastern edge of Prairie Divide. The present climate is not rigorous enough for the formation of

such a topographic feature at this altitude. The complete vegetative cover on the walls and floor of the nivation cirque indicates that the formative process is no longer active.

Ventifacts scattered upon the surface of Prairie Divide are the result of steady winds blowing across the high unprotected flat from the ice-choked mountains to the west and southwest. Where erosion has removed the mantle of glacial debris and exposed the granitic bedrock west of Prairie Divide, numerous small undrained depressions containing shallow ponds have been developed. Remnants of terraces which surround these ponds represent higher stands of the water level during pluvial periods related to the later Wisconsin glaciation. Ventifacts scattered around the borders of these lakes indicate the effectiveness of the periglacial winds. At present the ponds contain only a few inches of water; in dry seasons they may be completely dry. Residual boulders in the beds of the ponds have rinds of calcareous material precipitated during periods of drought.

Following the "Canyon Cutting Cycle," ice tongues of the Wisconsin substages moved from the high summit areas of accumulation down the narrow inner canyons of the major streams.

WISCONSIN GLACIATION

General statement.—There is no evidence of glaciation in the Cache la Poudre Valley up to an elevation of about 7500 feet (Figs. 2, 3). In general the valley is narrow and V-shaped, with little flat land along the stream, except for the many remnants of gravel terraces. Although King (1878, p. 467) reports a glacial moraine at an elevation of 6500 feet, the writer could find no evidence for it. The lowest moraine in the canyon occurs immediately below Home post office. For a short distance below this moraine, the valley, although steep-walled, has a U-shaped profile similar to that of the upper glaciated valley above the moraine. No till was found below the moraine at Home, and at first sight the terminal moraine appeared to mark the maximum advance of the glacier.

Twin Lakes and Home substages (Wisconsin I and II).—A careful study of the terrace remnants of the Cache la Poudre Valley and their correlation (Bryan and Ray, 1940) indicates a terrace above the outwash of the glacier that developed the Home moraine. This high terrace, the character of the valley, the position of certain erratics above the margin of the ice of the Home moraine, abandoned channelways, to be described later, and corroborative evidence from other regions lead to the belief that in the Cache la Poudre Valley the first Wisconsin substage was present, but almost all evidence of its existence has been removed, and its presence is only indirectly suggested. This pre-Home or Twin Lakes substage of

the Cache la Poudre Valley is correlative with the earliest moraines of the Leadville region (Fig. 1), from which it takes its name, and with the Durango glaciation of the San Juan Mountains (Atwood and Mather, 1932). The writer believes it is correlative with the Iowan glaciation of the central west, which Kay (1931a and b) and Leighton (1931) consider

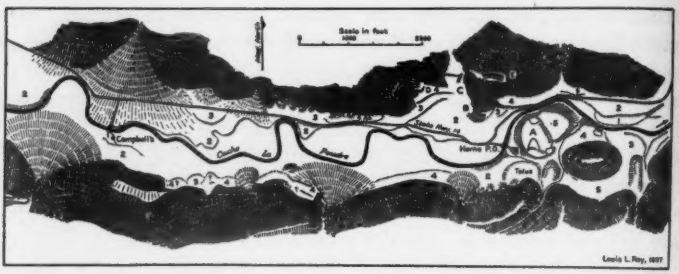

FIGURE 2.—*Sketch map of the Home moraine and Cache la Poudre valley*

(A) Home moraine; (B, C, D, and E) lateral stream channels cut in bedrock; high lateral moraine in lower left. Terraces correlated with later glacial substages are shown diagrammatically.

the earliest Wisconsin advance of the continental glacier. For purposes of nomenclature, this substage will in this paper be called the Twin Lakes substage or Wisconsin I because of its excellent development at Twin Lakes, Colorado (Davis, 1905; Westgate, 1905; Capps, 1909).

At Home post office a terminal moraine (Pl. 1, fig. 2) stretches almost completely across the Cache la Poudre Canyon. This moraine, at an elevation slightly above 7500 feet, rises to a maximum height of 135 feet above the present river level. In the valley above the moraine are scattered roche moutonnée surfaces, erratics, hanging tributary valleys, lateral moraines, and other features typical of glaciated valleys.

The Home moraine, from which the second Wisconsin substage takes its name (Ray, 1938; Bryan and Ray, 1940), is a smooth hillock. Most of the moraine rises about 100 feet above the stream, where it forms a broad flat, approximately 300 feet in width, normal to the axis of the valley, and 450 feet in breadth. Its general outline is that of an equilateral triangle, one of whose sides is normal to the axis of the valley (Fig. 2). From above, few boulders can be seen scattered across the surface of the moraine; most of them are concentrated in a small ridge along the frontal margin. This boulder ridge was probably formed by a concentration of debris dropped from the frontal cornice of the glacier. The few large boulders scattered on the surface of the moraine are of pre-Cambrian granite. They are split and wind polished but lack a decomposition rind or glaciated surface. No good till exposures can be

FIGURE 1. PRAIRIE DIVIDE
View of surface showing smooth gravel-capped till plain.

FIGURE 2. HOME MORAINE, CACHE LA POUDRE CANYON
Viewed from upstream side. River flows in front of moraine, from right to left and between moraine and valley wall. This is the moraine from which the Wisconsin II substage derives its name.

FIGURE 3. WISCONSIN III MORAINE IN EVANS GULCH
Near Leadville, Colorado.

PRAIRIE DIVIDE AND MORAINES

FIGURE 1. MORAINE ADJACENT TO TERMINUS OF SPRAGUE GLACIER, ROCKY MOUNTAIN NATIONAL PARK
It is from this moraine that the last glacial substage, the Wisconsin V, derives its name.

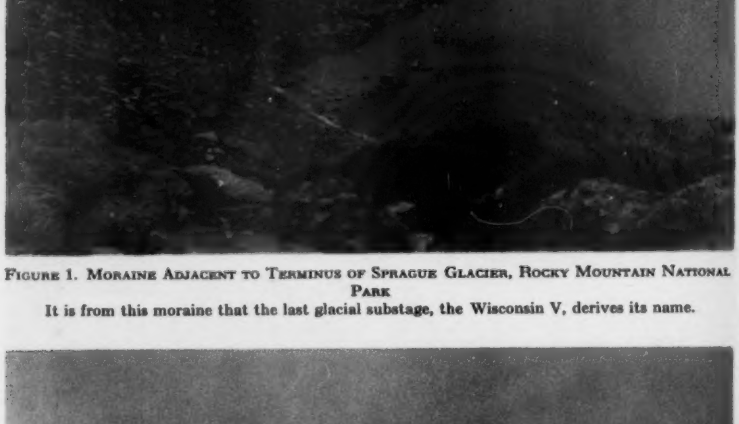

FIGURE 2. GLACIAL MORAINE OF THE WISCONSIN IV SUBSTAGE
Near the southern end of the Snowy Range, Medicine Bow Mountains.

MORAINES

seen on this moraine. A pit about 5 feet deep showed that the boulders near the surface were slightly iron stained. No highly decomposed boulders were found.

The subdued character of this moraine is not primarily the result of weathering or erosion but of a mantle of windblown sand which effectively conceals most of the underlying glacial material. Resting against the upstream side of the moraine are low dunes of finely cross bedded sand, remnants of vast quantities swept down the valley when the glacier receded. Blackwelder (1929) has described and illustrated similar conditions related to glaciation in the Sierra Nevada of California. Shifting sand dunes have also been observed by the writer in front of Alaskan valley glaciers (Ray, 1935, p. 312).

Glacial boulders on the Home moraine show sand scouring, grooving, and polishing. Some are well facetted, and one, too large to be removed, had braided, wind-scoured grooves almost an inch deep. Roches moutonnées and projecting portions of the valley wall adjacent to the Home moraine exhibit scoured and polished surfaces in places protected from present-day weathering.

In reconstructing the appearance of the ice as it must have been during its maximum advance at Home, the thickness and slope of the surface should be determined. This cannot be done precisely, for rocks of the valley walls are easily weathered and have been modified since they were scoured by the glacier. Pinnacles have been developed in the south valley wall, immediately upstream from the moraine. However, 2 miles upstream a high lateral moraine stretches partially across a small tributary valley (Fig. 2) about 850 feet above the present stream level. If this is a product of the Home substage, as the writer believes, it indicates that the ice surface was at least 715 feet above the top of the terminal moraine at Home and that the minimum slope of the ice surface was 3° 45′ for the first 2 miles. Above the projected ice surface of the Home substage and on the south valley wall adjacent to the moraine a few large erratic boulders are believed to have been deposited by the pre-Home or Twin Lakes substage of glaciation.

Abandoned stream channels immediately upstream from the Home moraine bear on the history of the glaciation. The lowest are the result of drainage along the margins of the glaciers of the Twin Lakes and Home substages (Fig. 2, B, D, and E), when streams of melt water were active agents of erosion. The highest of these abandoned channels (Fig. 2, D and E) due directly to glaciation is more than 250 feet above the river on the northern valley wall and appears to be much older than the lower channel (Fig. 2, B). This highest channelway is seemingly related genetically to the Twin Lakes substage. Its walls have entirely lost their

water-rounded surface, and small talus piles have filled the floor of the channelway. The lower channelway, about 60 feet above the river, was developed when the ice of the Home substage was shrinking. Melt waters from the surface of the ice flowed to and down the lateral margins, cutting across the low bedrock spurs. A long period need not be assumed for the excavation of such channels, for similar ones have been observed by the writer in Alaska, where they have formed rapidly along the lateral margins of retreating valley glaciers (Ray, 1935, p. 304-307).

The character of the lowest channelway indicates a greater degree of youthfulness than that of the older talus-filled and higher channelway. In this lowest box-canyon, scarcely more than 15 feet wide and 25 feet deep, the walls are wind polished, smoothed by water action, and show incipient potholes. Removal of the thin veneer of drift, which partially covers the roche moutonnée in which the channelway is cut, exposes a well-striated and polished fresh surface, which, like the small canyon, gives the impression of youthfulness wholly lacking in the higher channelways.

A quarter of a mile from the moraine on the upstream side, another small, deep channel has been cut into the north valley wall normal to the axis of the valley (Fig. 2, C). This channel was formed by the combined action of the runoff waters from a small drainage basin, whose area is slightly more than a square mile, and the surface and lateral runoff from the glacier of the Home substage. Its formation was thus contemporaneous with the one previously described (Fig. 2, B).

Corral Creek substage (Wisconsin III).—The higher reaches of the valley of the Cache la Poudre River and its tributaries are choked by moraines at elevations from about 9100 to 10,100 feet or from 1500 to 2500 feet above the moraine of the Home substage (Fig. 3).

The moraines of the Corral Creek substage comprise a group with more or less accordant altitude and were formed when ice tongues pushed forward 4 to 5 miles from the cirques with less vigor than during the preceding Home substage (Ray, 1938; Bryan and Ray, 1940). Unlike the Home moraine, those of the Corral Creek substage are generally not a single barrier of debris stretched across the valleys but a morainic complex with a series of recessionals, as at Chambers Lake, Trap Creek, and the Cache la Poudre River at Chapin Creek (Fig. 3). The Corral Creek substage, named for the excellent moraine in the valley of Corral Creek, in many places represents a series of minor fluctuations, slight advances, and recessions.

The materials of this substage are well exposed in the Chambers Lake region (Fig. 3), where road cuts have been made through the moraines. Here the till and boulder clay are fresh and unweathered. The fine-

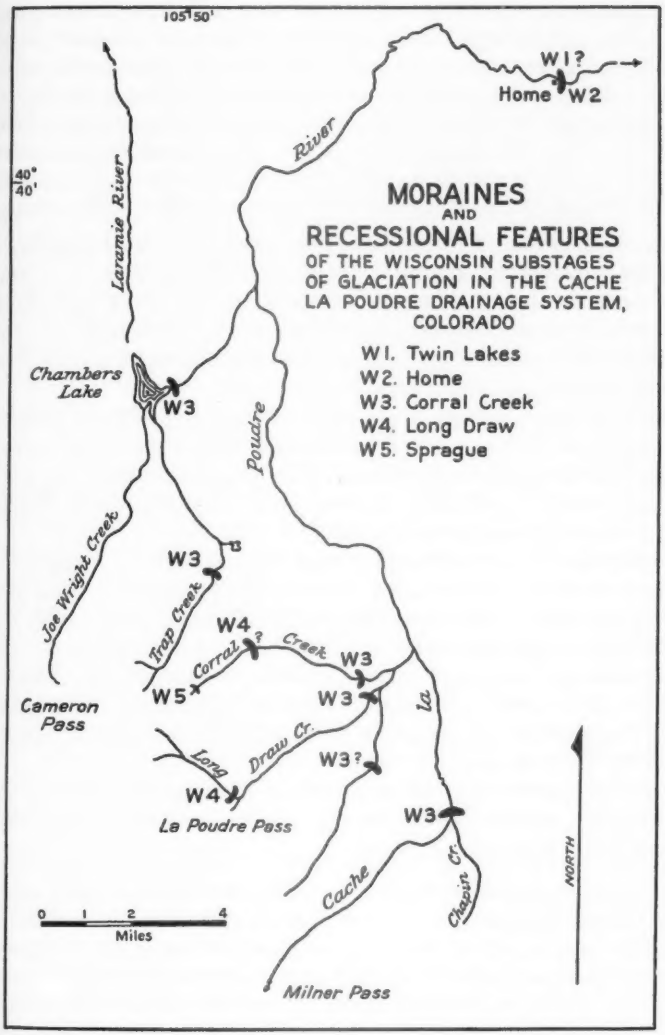

FIGURE 3.—*Sketch map showing location of moraines in the Cache la Poudre drainage system*

grained matrix of the till is a light blue gray and the boulders, except for a few, are striated, fresh, and unweathered. Occasional deeply weathered boulders are derived from debris of earlier glacial substages.

The validity of this third Wisconsin substage, the Corral Creek, rests upon four criteria which will be discussed later: (1) Topographic position of the moraines, (2) character of the till, (3) regional development of moraines similar in character, and (4) terrace deposits of the Cache la Poudre Canyon, which correspond to the outwash deposits of this substage.

Long Draw substage (*Wisconsin IV*).—In a small valley tributary to Long Draw (Sec. 17, T. 6 N., R. 75 W.), approximately 4 miles upstream from the moraines of the Corral Creek substage (Fig. 3), a low moraine was formed by a small ice tongue pushing about 2½ miles from the cirque headwall. This moraine has been taken as the type for the Long Draw substage (Bryan and Ray, 1940).

The Long Draw moraine is composed of fresh, unweathered boulders, dominantly of Tertiary volcanic rocks and subordinate pre-Cambrian granite. Sand in large quantities partially mantles the moraine, giving it a low, rounded appearance, in contrast to what one would normally expect in such a young morainic deposit.

In the valley of Corral Creek no moraine was found to correspond with that of Long Draw. However, its place is taken by a pitted outwash plain slightly more than 2 miles upstream from the Corral Creek moraine (Fig. 3) and about 2 miles below the cirque headwall, at a point where the valley gradient becomes less steep. Local conditions, presumably the width and gradient of the valley floor, were such that no terminal moraine was built but an outwash plain developed in front of the ice terminus to represent the Long Draw substage.

The regional study here reported confirms the belief that this is a definite substage, separate and distinct from the earlier Corral Creek, and not a local halt in the glacial recession.

Sprague substage (*Wisconsin V*).—Within the Corral Creek cirque a ridge of rock debris from a distance appears to be a moraine. Close inspection, however, shows that it is composed of angular rock fragments showing no signs of glacial origin. A feature of this type has been termed by Bryan (1934) a protalus rampart. It owes its origin to the intensive frost action which mechanically disintegrates the bedrock of the cirque headwall so that angular blocks of detritus roll across snow banks occupying the cirque, building a debris barrier at the foot of the snowbank. This feature has been termed a nivation ridge by Behre (1933a).

In order to build a rampart such as that of the Corral Creek cirque

more snow is needed than is now supplied. No accretion to this rampart is being made at present, for lichens are growing undisturbed on the angular blocks, and occasional small shrubs are growing between them. The fosse between the protalus rampart and the cirque headwall is being filled by talus material from the latter. Climatic factors which operated to form the protalus rampart are no longer active.

In areas where bedrock is less favorable to rock disintegration than the Tertiary volcanic flows in the region of the Corral Creek cirque, or where the position of the cirque is such that the exposure is not conducive to the formation of large snowbanks, protalus ramparts are not developed and talus accumulations may be small.

The cool period when the protalus rampart of the Corral Creek cirque was developed was not severe enough to produce a glacier tongue but was enough to fill the cirque with a semipermanent snow mass. This climatic condition has been only slightly modified, for small patches of snow lay in this cirque in early September 1936 when it was first visited by the writer. In years of heavy snowfall or unusual coolness snowbanks may last from one season to another. In Rocky Mountain National Park, to be described later, a moraine in front of Sprague Glacier formed contemporaneously with the protalus ramparts elsewhere. For this reason the protalus rampart, or Wisconsin V substage, of the earlier reports (Ray, 1939; Bryan and Ray, 1940) is here renamed the Sprague substage.

The Sprague substage is the last event definitely recorded in the cirques and may have occurred after the time usually designated as the end of the glacial period. Because of the indefiniteness of the dating of the close of the glacial period in North America, this last period of climatic refrigeration is called Wisconsin V. Future study may show that this substage should be assigned to postglacial time, when that time is more adequately defined.

GLACIATED PASSES AT HEADWATERS OF CACHE LA POUDRE DRAINAGE SYSTEM

Three passes at the headwaters of the Cache la Poudre drainage were examined—La Poudre and Milner, both on the Continental Divide, and Cameron, between the headwaters of the Cache la Poudre and Laramie River drainage (Rocky Mountain National Park topographic map). These passes are low, broad cols, at elevations of 10,194, 10,759, and 10,200 plus feet respectively. At each, the grade to the crest from the eastern side is gentle, but across the pass on the western side it steepens (Fig. 4). No explanation of the differences in gradient will be attempted here. These passes have been glaciated, but not since the Home substage.

Stratified gravel was found at the crest of Cameron and La Poudre

passes. Professor O. P. Pennock of the Colorado State College, who a[illegible] in the construction of an irrigation ditch across La Poudre Pass a[illegible] 20 years ago, told the writer that there were peat deposits about 20 feet thick at the crest of the pass. At Milner Pass two small lakes are perched

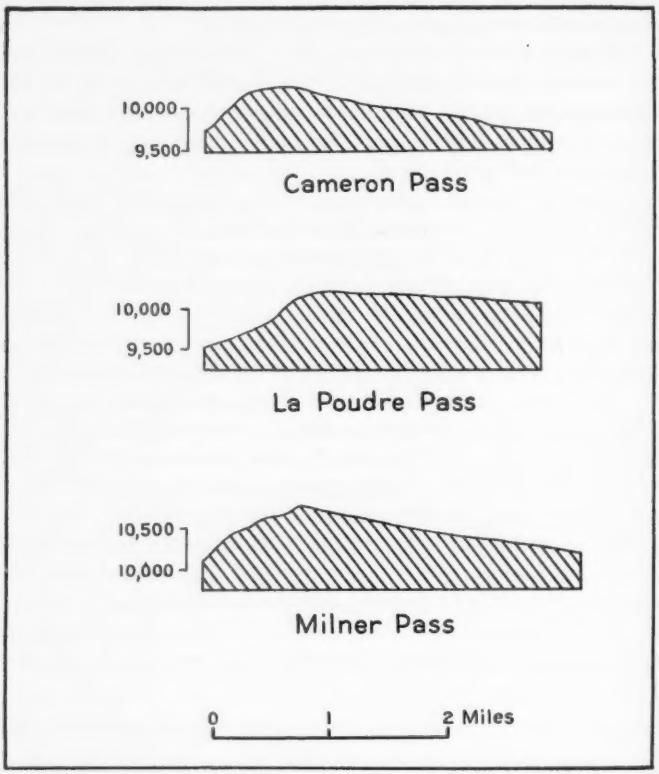

FIGURE 4.—*Profiles of passes at the headwaters of the Cache la Poudre drainage*

on the crest of the divide. These peculiar relationships have not been found elsewhere in the Colorado Front Range.

In the gently eastward sloping valleys below these passes glacial till and scattered patches of outwash gravel indicate the occupation of the lower valleys by glaciers of the Corral Creek substage. There are no cirques which could shed ice directly on the passes. Thus, when these valleys were ice-filled in their lower portions, a natural dam was provided

to impound water to the level of the crest of the passes. Glacial debris was deposited in these lakes, and stratified gravel was laid to the crest of the passes. With the retreat of the glaciers the lower valleys became ice free, and drainage resumed its normal course. The ice tongues of the Long Draw substage were not large enough to dam the main valleys and form lakes.

SUMMARY OF GLACIATION OF CACHE LA POUDRE DRAINAGE BASIN

There is evidence in the Cache la Poudre drainage basin of two major stages of glaciation—the pre-Wisconsin Prairie Divide stage and the Wisconsin. The latter has been divided into five substages. Four lines of reasoning lead to the belief that the substages are separate and distinct, not simply retreatal phases of the Wisconsin glaciation:

(1) The separate and distinct morainal systems have a regional rather than a local expression.

(2) The amount of weathering on the moraines varies. Differences in weathering between the Home and Corral Creek substages are distinctive although not excessive.

(3) The moraines of each substage are the source points of outwash plains which are preserved in the valley of the Cache la Poudre River as terrace remnants. These remnants are separate and distinct and may be traced from the moraines down the canyon and into the drainage of the South Platte River to the east (Bryan and Ray, 1940).

(4) Evidence of the development of a soil zone between the Wisconsin I and II substages is reported in the following section.

ROCKY MOUNTAIN NATIONAL PARK AND ADJACENT REGIONS

GENERAL STATEMENT

No portion of the Southern Rocky Mountains has attracted the attention of the glacial geologist more than the Rocky Mountain National Park region. Numerous popular as well as technical papers have dealt with the glaciation of these mountains. A generalized and concise picture of the glaciation may be obtained from Lee's semipopular treatise (1917). Glacial features shown on the Rocky Mountain National Park topographic map are familiar to all students of geology.

Glaciers in this region moved from the cirques down to elevations of approximately 8000 feet. The more important of these glaciers have been named by Lee (1917, p. 31). The five largest, here discussed, are: Fall River, Thompson, Bartholf, Wild Basin, and Monarch Valley (Fig. 5).

Lee (1917, p. 31) pointed out that there are probably two stages of

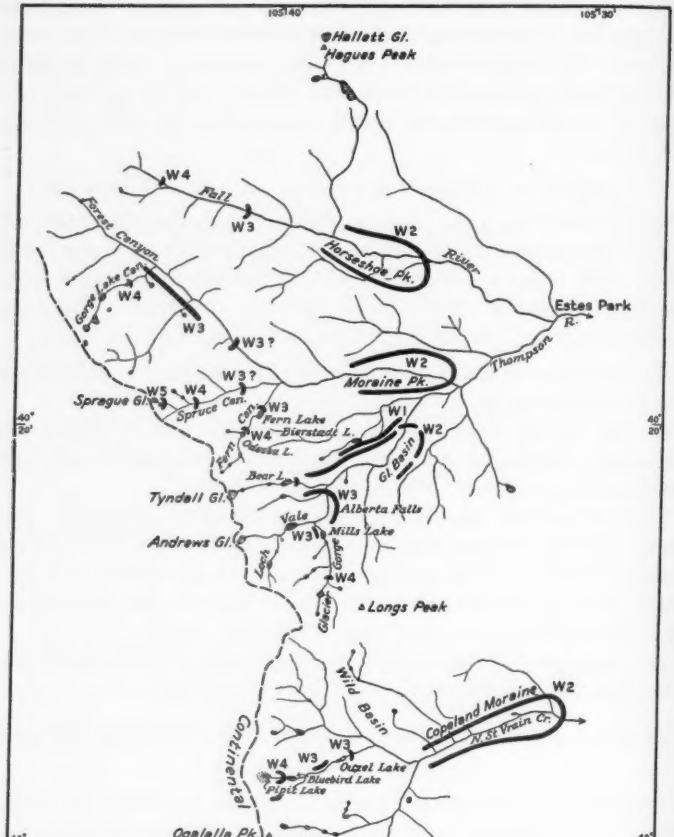

FIGURE 5.—*Sketch map of Rocky Mountain National Park region*
Showing location of moraines examined.

glaciation represented—an older pre-Wisconsin and a younger or Wisconsin. Of the earlier glaciation, he says (1917, p. 31),

> "Little need be said . . . of the older stage, for proofs of its existence are not entirely conclusive. Apparently such broad open valleys as Estes Park and Tahosa Valley were shaped by glacial ice . . . but they lie outside of the area affected by the glaciers of the Wisconsin stage."

The present writer, however, believes that Estes Park and Tahosa Valley are not the results of glaciation, for which no evidence was seen, but are explained by normal weathering and erosive processes. They were probably developed in pre-Wisconsin time, perhaps during the "Canyon Cutting Cycle," between the Prairie Divide and the Wisconsin stages. The Wisconsin I substage is believed to have occupied the present inner valleys—a condition analogous to that of the Cache la Poudre Canyon.

Fuller (1923) has advanced the theory of a pre-Wisconsin glaciation in this region, reporting till in the valley of the Big Thompson River at the eastern end of Mont Rose Valley, at an elevation of about 5600 feet. The present writer could find no evidence of this glacial stage within the inner valley of the Big Thompson River. The criteria used by Fuller in establishing this glacial stage are not deemed critical enough to warrant belief in the existence of such a pre-Wisconsin ice tongue within the present stream valley.

Glaciers of the Wisconsin stage in this region did not descend below elevations of about 8000 feet. Near Estes Park they terminated in the broad basins of Horseshoe and Moraine parks (Fig. 5 and Rocky Mountain National Park topographic map), where they deployed from their confining valleys into the broad parklike areas and expanded into great terminal bulbs. A large surface of melting was thus given the terminal ice so that, although the summit area of accumulation was greater for these glaciers than that of the Cache la Poudre glacier, the terminal ice was dissipated at higher elevations than the ice tongue confined within the narrow Cache la Poudre Canyon.

The substages of glaciation in this region are similar to those previously described from the Cache la Poudre Canyon. Morainal deposits of the Wisconsin II substage are most easily recognized. The Wisconsin I substage can be inferred only from scattered field evidence. A more detailed study may reveal that the morainal deposits mantle older deposits and that in some places there may be older deposits in front of and adjacent to the till of the Wisconsin II substage. Near Ward, Colorado, about 10 miles south of Rocky Mountain National Park, Wahlstrom (1939) reports the finding of an old soil zone below a younger till, which caps the great lateral moraine on the north side of North Boulder Canyon, about 7 miles below Arapaho Glacier. Here, a great lateral moraine was built

by ice of the Wisconsin I substage, and was later mantled by till of the Wisconsin II substage. The old soil horizon offers confirmatory evidence for the belief in a relatively long interstadial period between the first two Wisconsin substages. Similar covering of the Wisconsin I till by Wisconsin II till may have taken place on the great lateral moraines of Rocky Mountain National Park.

Reconnaissance work in the region under discussion consisted of an examination of: (1) Fall River Valley, (2) Spruce Canyon and its continuation as Forest and the Thompson River Canyon, (3) Glacier Gorge and Glacier Basin, (4) North St. Vrain Valley, (5) Middle St. Vrain Valley, and (6) Monarch Valley, on the western slope of the mountains, across the Continental Divide. The Wisconsin substages of these valleys are discussed in this order.

FALL RIVER GLACIER

Ice of the Wisconsin II substage filled the valley of Fall River with an ice tongue almost 10 miles long (Fig. 5). This glacier terminated at an elevation of about 8200 feet in the Horseshoe Park basin, leaving a poorly defined terminal moraine. Great lateral moraines, rising to a maximum of about 900 feet above the floor of Horseshoe Park, are the most prominent glacial feature in this valley.

Approximately 5 miles up the valley, in sections 9 and 10, T. 5 N., R. 74 W., a mass of jumbled morainal material fills the valley of Fall River, between elevations of about 8900 and 9300 feet. This is interpreted as morainal debris of the Wisconsin III substage. Grooved and striated bedrock surfaces are related to this debris. A short distance upstream post-Wisconsin III erosion has permitted Fall River to cut a box canyon about 20 feet deep into the bedrock of the valley floor.

At an elevation of about 10,700 feet (Fig. 5) is a low moraine of fresh blue-gray boulder till. This is correlative with the Wisconsin IV substage and is the product of a small ice tongue which pushed only about 1½ miles from the cirque headwall. The gently sloping vegetation-covered headwall of the cirque immediately below Fall River Lodge shows no protalus rampart or other evidence of the Wisconsin V substage.

Thus, in the valley of Fall River definite features can be assigned to the Wisconsin substages II, III, and IV. The Wisconsin I is still unidentified; the Wisconsin V, because of local conditions, was never developed.

THOMPSON GLACIER

One of the largest glaciers in the northern Colorado Front Range moved about 15 miles (Fig. 5) down the valley of the Thompson River, from the head of Forest Canyon to Moraine Park. It was fed from the large cirques west of Forest Canyon, nestled beneath the high crests of the

Continental Divide. (*See* Rocky Mountain National Park topographic map.) Today, all these areas of accumulation show extreme glacial modification, and one, Spruce Canyon, supports a small permanent ice mass known as Sprague Glacier. Spruce Canyon and the adjacent Fern Canyon were visited by the writer, who followed them to the terminus of the major ice tongue in Moraine Park.

During the reconnaissance examination of the moraines of Moraine Park, no definite evidence of the Wisconsin I substage was observed. However, it is possible that the bulk of the great lateral moraines which almost encircle the park were formed during the Wisconsin I substage and have been mantled by debris of the Wisconsin II substage. The lateral moraine on the south side of the park rises almost 800 feet above the stream, and its even crest makes it one of the most prominent glacial features of the entire area. The floor of Moraine Park supports numerous gravel terrace remnants, one of which rises consistently about 60 feet above the stream. This terrace level was not observed above the entrance to Forest Canyon, where dense vegetation cloaks the valley bottom. The glacial outwash of which this terrace is built is correlated by the writer with the Wisconsin III substage, whose ice tongue did not extend to the mouth of Forest Canyon.

There are no traces of glacial moraines between those of Moraine Park and the mouth of Spruce Canyon. Ice tongues of the Wisconsin III substage did not move from the hanging tributary valleys of Fern and Spruce canyons into the main valley of the Thompson River but descended to an elevation of only about 9500 feet in Fern Canyon, where a terminal moraine was built, which now holds in Fern Lake. In Spruce Canyon a moraine belonging to this substage was not observed because of the dense vegetation, although it is probably at an elevation between 9000 and 9500 feet.

Forest Canyon was not traversed above the mouth of Spruce Canyon. The writer believes, however, that during the Wisconsin III substage tributary glaciers from the cirques to the west must have coalesced and moved downstream to some point between the mouths of Hayden and Spruce canyons. This ice tongue is believed to have built the lateral moraine, about 500 feet high, which dams a small lake at the mouth of Gorge Lake Canyon. (*See* Figure 5 and Rocky Mountain National Park topographic map.)

The Wisconsin IV substage is well represented. In Fern Canyon it is a small glacial moraine at an altitude of 10,000 feet, which holds in the waters of Odessa Lake. In Spruce Canyon a similar moraine occurs at an altitude of 10,100 feet. This substage is represented in Gorge Lake Canyon by a small moraine which holds in a lake at an altitude of

10,300+ feet. Glaciers of this substage advanced scarcely more than 2 miles from the cirque headwalls.

Heavy talus in some cirques offers slight evidence for a Wisconsin V substage. Most important proof for its existence is found immediately below Sprague Glacier at the head of Spruce Canyon. Here, a tiny ice mass, dwindling in a small, sheltered cirque, is perched almost 1000 feet above the floor of Spruce Canyon. A morainic ridge (Pl. 2, fig. 1) at the lip of this cirque is assigned to the Wisconsin V substage; from it the substage has derived the name Sprague. Sprague Glacier terminates in a small lake about a quarter of a mile behind the terminal moraine, which lies at an elevation of about 11,800 feet. This small Wisconsin V moraine cannot be confused with that of the Long Draw moraine, which lies at an elevation about 1700 feet lower. The Sprague moraine represents the last important period of ice advance.

The valley of the Thompson River presents evidence of four substages of Wisconsin glaciation, for all of which there are moraines, more or less accordant in the various tributary valleys. No tangible evidence of the existence of the Wisconsin I substage was observed by the writer. Its presence, however, may be inferred from studies of near-by regions.

BARTHOLF GLACIER

Ice tongues pushing from the cirques below the Continental Divide through Glacier Gorge, Loch Vale, and smaller valleys once coalesced to form Bartholf Glacier (Fig. 5) described by Lee (1917, p. 33). This glacier, about 8 miles long, terminated in Glacier Basin at an elevation of slightly more than 8000 feet. Along the north wall of Glacier Basin the ice built a great lateral-terminal moraine, on whose crest is perched Bierstadt Lake (Fig. 5 and Rocky Mountain National Park topographic map.) Although the evidence is none too definite, the writer believes that the greater part of this moraine was formed by the Wisconsin I ice advance and not by the Wisconsin II which has in the past been held to mark the greatest extension of the Wisconsin ice. Along the side of Bierstadt moraine, small and more or less accordant flats indicate the upper limits of later glacial debris deposited on the sides of the older moraine of the Wisconsin I substage. From this topographic evidence, supported by the regional study, the writer believes that the first two Wisconsin substages occupied Glacier Basin, developing a complex lateral moraine.

The Wisconsin III substage is represented by a small morainic loop below Alberta Falls (Fig. 5 and Rocky Mountain National Park topographic map) built by small coalescing glaciers which moved from Glacier Gorge and Loch Vale. Similar glaciers moved from cirques now occupied

by Tyndall Glacier and small permanent snow masses to build a small moraine below Bear Lake. Ice of this substage descended to an elevation of about 9300 feet. A small lateral moraine below Mills Lake in Glacier Gorge belongs to the Wisconsin III substage.

The Wisconsin IV substage of the Bartholf Glacier is poorly represented and has been observed only in Glacier Gorge, where a low moraine at an elevation of 10,300 feet stretches across the valley floor. Because of the massiveness of the bedrock, little talus material has accumulated in Glacier Gorge, where there is no evidence of the Wisconsin V substage.

Tyndall, Andrews, and Taylor glaciers, which were not visited by the writer, lie in sheltered cirques in the area of accumulation of Bartholf Glacier. In the valleys immediately below these glaciers there may be small moraines belonging to the Wisconsin IV and V substages.

WILD BASIN GLACIER

During the earliest Wisconsin substage, the North St. Vrain valley was filled by a glacier about 9 miles long, fed by numerous cirques along the Continental Divide, from a point near Longs Peak to Ogalalla Peak. (Fig. 5 and Rocky Mountain National Park topographic map.) During the Wisconsin II substage this glacier extended to a point slightly below 8300 feet, building great lateral moraines in the lower portion of the valley. These, rising about 700 feet above the floor of the present valley, are similar to those of the Fall River, Thompson, and Bartholf glaciers. The moraine on the north valley wall is known as the Copeland Moraine. The writer believes that these great moraines, like those of Horseshoe and Moraine parks, are probably the combined product of the Wisconsin I and II substages of glaciation.

The valley of North St. Vrain Creek was traversed to Pipit Lake, a tarn below the Continental Divide, at an elevation of about 11,400 feet (Fig. 5). No moraines were observed in the heavy timber of the valley from the terminal moraine of the Wisconsin II glaciation to the mouth of Ouzel Creek, at an elevation of about 9500 feet. Here scattered morainic debris, believed to be the product of the Wisconsin III substage, holds in Ouzel Lake at an elevation of more than 9800 feet.

About a mile upstream from Ouzel Lake, a small lateral moraine of the Wisconsin III substage holds Chickadee Lake in the fosse between the moraine crest and the valley wall. The lake is unnamed on the topographic map.

Above the great rock step and above the tarn, Bluebird Lake, is a moraine at an elevation of 11,200 to 11,300 feet. It is typically fresh in appearance and is the product of the Wisconsin IV substage, when glaciers

moved about a mile from the cirque headwall. This moraine holds in the waters of Pipit Lake.

Large quantities of talus material have accumulated in the Pipit Lake cirque, but no protalus rampart has been developed. It is possible, however, that much of the existing talus was produced during the Wisconsin V substage, when the climate was more rigorous than at present. The southern exposure of the cirque was unfavorable to the accumulation of great snowbanks on which the formation of protalus ramparts is dependent.

MIDDLE ST. VRAIN GLACIER

South of Rocky Mountain National Park (Fig. 5 and Rocky Mountain National Park topographic map) is the great valley of the Middle St. Vrain Creek, at whose head nestle small cirques containing the St. Vrain Glaciers. At the time of the greatest extension of the glacier occupying the valley of Middle St. Vrain Creek, the Wisconsin II ice tongue moved as far downstream as Peaceful Valley, to an elevation below 8500 feet. The character and extent of the Wisconsin I glacier are unknown.

South of and above the present valley of Middle St. Vrain Creek is a great flat, covered by ancient glacial debris, on which are Stapp Lake and Beaver Reservoir. (*See* Rocky Mountain National Park topographic map.) Glacial debris composing this flat is believed to belong to a pre-Wisconsin stage, deposited before the cutting of the present inner canyon of Middle St. Vrain Creek. No detailed examination was made of this debris, but it is tentatively suggested that it is the correlative of the Prairie Divide stage so well developed to the north.

The lateral moraines of the Wisconsin I and II substages are neither large nor conspicuous. The Wisconsin III substage is represented by small masses of hummocky morainal debris at an elevation of approximately 9700 feet (Fig. 5) and not by a conspicuous morainal complex, as in the Cache la Poudre Canyon.

No debris which could be interpreted as a terminal moraine occurs between the Wisconsin III moraine and an elevation of 10,800 feet. From this point, about a mile below the present permanent ice masses—the St. Vrain Glaciers—a jumbled mass of till and talus is scattered along the valley floor to a point immediately below the ice front, where the till is partly covered by talus. This till is correlated with the Wisconsin IV substage, and the talus with the climatic refrigeration of the Wisconsin V. Some till and talus are being added now to this jumbled mass. Milky water issuing from the glacier termini indicates present activity of the ice masses.

Although these glaciers were visited in late July 1938, they were still almost completely covered by snow of the previous winter, and only

scattered patches of ice granules with a clearly defined stratification were observed.

MONARCH VALLEY GLACIER

The writer spent parts of 2 days examining the lower moraines in the Monarch Valley (*See* Rocky Mountain National Park topographic map) reported by Ives (1938, p. 1052). Here, Wisconsin terminal moraines are well developed and free from a heavy vegetative cover. On the north valley wall a great lateral moraine rises almost 900 feet above the valley floor and is here assigned to the Wisconsin I substage. This has been called the Stillwater Stage by Ives, who believes it to be "not younger than Illinoian" (Ives, 1938, p. 1062). However, the morainic material composing the bulk of this moraine is no more weathered than that of the Durango (Atwood and Mather, 1932) and the Wisconsin I moraines in the Leadville region. Material of this first Wisconsin glaciation is only slightly more weathered than that of the Home moraine (Wisconsin II) and is very fresh and unweathered compared with the deeply decomposed till of the Prairie Divide and Cerro stages of glaciation.

Differences between the glacial debris which composes the three moraines of the River and Arapaho (double) stages of Ives, at an elevation of about 8100 feet, are so slight that the present writer, on the basis of the reconnaissance here reported, would place them all in the same Wisconsin II substage and correlate them with the Home moraine of the Cache la Poudre Valley and the prominent moraines on the eastern slope of the Colorado Front Range. The long interval separating the Wisconsin I and II substages, indicated by the soil zone reported by Wahlstrom (1939) and recorded topographically in Glacier Basin, marks the time between the formation of the great lateral moraine in the Monarch Valley and the younger Wisconsin II terminal moraines and low lateral moraines which lap on to the older Wisconsin I lateral moraine.

The complex system of terminal and lateral moraines reported by Ives (1938, Fig. 2) and named the River and Arapaho stages are held by the present writer to be simple products of a slightly oscillating ice front. The writer was not able, after field examination, to believe a subdivision of these morainic ridges into separate stages necessary, either on the basis of topographic expression or depth of weathering.

Five miles up the valley from the mouth of Arapaho Creek (Ives, 1938, Fig. 3) lies another morainic complex, termed by Ives the Monarch stage. This substage is here correlated with the Wisconsin III or Corral Creek moraines of the Cache la Poudre valley.

The writer did not examine the drainage basin of Arapaho Creek above the confluence with the East Fork. However, Ives reports that the valleys

are free from moraines to a point about a mile below the cirques. This is based on Ives' statement that

"glaciers were re-established in many of the high cirques, and advanced not more than a mile down the valleys" (1938, p. 1061).

This advance is correlated with the Wisconsin IV substage. The Wisconsin V substage seems to be represented by moraines which are reported by Ives to occur only a few hundred feet from the headwalls of many of the cirques where some of them form the dams of small lakes.

It seems, therefore, that on the western slope of the Colorado Front Range in the Rocky Mountain National Park region, Wisconsin substages are similar to those on the eastern slope. Table 1 gives a reinterpretation of the glacial substages of the Monarch Valley and correlates them with the other glaciated areas of the Southern Rocky Mountains.

TABLE 1.—*Correlation of glacial substages in the Monarch Valley (Ives) with the general chronology of the Southern Rocky Mountains*

Ives (1938)	Ray (1939)
Monarch Valley	Southern Rocky Mountains
Unnamed advance	Wisconsin V (Sprague)
Recent advance	Wisconsin IV (Long Draw)
Monarch (Late Wisconsin)	Wisconsin III (Corral Creek)
Arapaho (Middle Wisconsin) River (Early Wisconsin, Iowan)	Wisconsin II (Home)
Stillwater (not younger than Illinoian)	Wisconsin I (Twin Lakes)

MEDICINE BOW MOUNTAINS

In the summer of 1937 the writer first visited the Medicine Bow Mountains to examine the moraines of Libby Creek Valley described by W. W. Atwood, Jr. (1937). The simple glacial sequence reported by Atwood did not fit the more complex chronology determined for the Cache la Poudre drainage basin in 1936. It seemed that in these two regions, separated by so few miles (Fig. 1), there should be a close correlation between the glacial advances and retreats.

With this apparent lack of correlation in mind, the writer returned to the Medicine Bow Mountains in the summer of 1938 to re-examine the moraines in the valley of Libby Creek and in the cirque area immediately below the Snowy Range (Medicine Bow Quadrangle). The present interpretation of the glacial substages reported by Atwood and the identification of younger substages result from these brief examinations.

Atwood recognizes only two glacial stages, the earlier of which he terms "pre-Wisconsin." This glacial advance produced moraines which, in many of the valleys, lie immediately downstream from glacial moraines that have been generally accepted as Wisconsin. Atwood (1937, p. 121-122) has assigned these moraines to the pre-Wisconsin on the basis of: 1) mature, gently rolling topography, 2) lack of lakes and swamps, characteristic of young morainal deposits, and 3) more severe weathering of the boulders than is characteristic of the younger moraines. These criteria, however, do not seem sufficient to warrant an antiquity so great that the deposits must be considered pre-Wisconsin. In many areas in the Southern Rocky Mountains, moraines which may be assigned to the Wisconsin stage have a similar "mature" form. This topography may be the result of the initial topographic character and position of the morainal material, as at Home, Colorado, and in the vicinity of Leadville. In these areas, Wisconsin moraines are gently rolling and may present a "mature" aspect, especially when they are mantled by windblown sand, as at Home. In the Southern Rocky Mountains, undrained depressions are not generally associated with early Wisconsin moraines, except at La Jara Reservoir (Atwood and Mather, 1932, p. 124, Pl. 24A) and in the Huerfano valley.

The depth of weathering on the "pre-Wisconsin" moraine near Centennial (Atwood, 1937, Fig. 7) is difficult to determine because the greater part of the boulders are of pre-Cambrian quartzite. Without doubt this debris is considerably older than that of the moraines up the valley near Libby Lodge. While it is extremely hazardous to attempt a correlation of glacial deposits in the Rocky Mountains from one area to another solely on the basis of the depth of weathering it may be done if differences in weathering are great and due attention is paid to the character of the debris, local climatic conditions, and topographic position of the till. These offer so many variables that depth of weathering should be sparingly used and never considered as a single criterion.

Because of its topographic position and lack of extreme weathering which is characteristic of the definitely pre-Wisconsin till of Prairie Divide, the "pre-Wisconsin" moraines of Atwood are here assigned to the Wisconsin I substage. In the valley of the Cache la Poudre the correlative glacial substage is established only by indirect evidence. However, glacial deposits of this substage occur in many valleys of the Southern Rocky Mountains. The similarity in character and topography to the Durango till of the San Juan Mountains and its position slightly in advance of the moraines assigned to the Wisconsin II substage lead the writer to the conclusion that this earliest Wisconsin substage in the Medicine Bow Mountains is correlative with the Durango till of the San Juan

Mountains. Atwood and Mather (1932, p. 130) have suggested that the Durango may be "Iowan," but it is here referred to the Wisconsin I substage.

Immediately behind the moraines of the first Wisconsin substage is an area covered with glacial debris of a much fresher aspect. This material (Atwood, 1937, Fig. 3) appears to represent at least two morainic systems—one at the outer margin, which impinges on the Wisconsin I moraines, and one just below Libby Lodge, at an elevation of about 8500 feet. The freshness of this drift, its topographic position, and relation to other moraines place it in the Wisconsin II substage or the correlative of the Home moraine in the Cache la Poudre Valley. It is possible that here the Wisconsin II substage is represented by a double morainic system.

Near the University of Wyoming Summer Camp is a well-defined morainic system at an elevation of approximately 10,000 feet. The till is almost wholly unweathered. The complex moraine surface presents a much younger topographic aspect than that in the lower valley, near Libby Lodge. It seems to bear the same relation to the lower moraines that the Corral Creek moraines of the Cache la Poudre region bear to the Home moraine, and it is for this reason assigned to the Wisconsin III substage.

Stretching for almost 2 miles from the base of the cirques nestled along the eastern face of the Snowy Range is an extremely youthful morainic topography, here assigned to the Wisconsin IV substage, the correlative of the Long Draw in the Cache la Poudre region. The hummocky, bouldery topography, numerous undrained depressions containing small lakes and swamps, and the lack of a dense forest cover increase the youthful aspect of the deposits. At an elevation of about 10,500 feet, rough polygonboden are the result of solifluction. Peat deposits in the swamps of this area are thin. Dr. P. B. Sears (personal communication) has found no evidence of peat deposits more than about 1500 years old. This leads to the belief that only in the last 1500 years has the climate been mild enough to support vegetation. The climate may have been cold enough to prevent the development of a strong vegetative cover but not moist enough to maintain permanent snow banks or small glaciers.

Protalus ramparts were not seen near the cirque headwalls, although there are large talus accumulations (Pl. 2, fig. 2). This may be explained by the smaller amount of precipitation and greater resistance to frost action of the massive bedrock of the Snowy Range than that of the highly fractured volcanics of the Corral Creek cirque during the Wisconsin V substage.

In summary, at least four Wisconsin glacial substages are represented by moraines in the Libby Creek area. These are correlated in Table 2.

TABLE 2.—*Correlation of the Wisconsin glaciation of the Medicine Bow Mountains (Atwood) with the Cache la Poudre moraines and the general chronology of the Southern Rocky Mountains*

Elevation of moraine (feet)	Atwood correlation	Ray correlation	Cache la Poudre moraines	Elevations (feet)
—?—	Wisconsin	Wisconsin V	Protalus	11,500
10,500	un-	Wisconsin IV	Long Draw	10,200
10,000	differenti-	Wisconsin III	Corral Creek	9,100–10,100
8,500	ated	Wisconsin II	Home	7,600
8,100	Pre-Wisconsin	Wisconsin I	Twin Lakes	7,500 ?

MOUNT EVANS AND VICINITY

During the Pleistocene the high mountainous regions west of Denver (Fig. 1) contained numerous large glaciers. In many places morainic deposits choke the valley bottoms at elevations well below 9000 feet.

Glacial deposits in the lower Clear Creek valley and its major tributary, Chicago Creek, were examined. Clear Creek has its source in the great cirques on the eastern side of the Continental Divide at Loveland and Berthoud passes (Central City, Fraser, Georgetown, and Montezuma topographic quadrangles). Chicago Creek drains the great cirques on the northern and western slopes of Mount Evans (elevation 14,260 feet).

Ball (1908) recognizes two important glacial stages in the Georgetown Quadrangle. These are designated the Earlier and Later Glacial epochs. The deposits of these stages are shown on Plates 2, 4, 5, and 80 of the United States Geological Survey Professional Paper 63 (Spurr and Garrey, 1908). Later substages are not designated. The present discussion is largely a reinterpretation of the deposits recognized in Professional Paper 63, together with added material demonstrating the existence of later glacial advances.

Ball (1908) found that the glaciers of the earliest stage occupied valleys deeply incised in an old mountain upland. Excavation of these valleys is presumably the result of the same period of incision which has been termed the "Canyon Cutting Cycle." No traces of glacial deposits older than this "Canyon Cutting Cycle" have been found in the Georgetown region, although presumably it was glaciated during the pre-Wisconsin stage, called the Prairie Divide in the northern Front Range.

In the valley of Clear Creek, as shown by Ball (1908, Pl. 2), the glaciers extended down to an elevation of approximately 7900 feet at a point about 3 miles upstream from the town of Idaho Springs, where the earliest

glacier had its terminus. Deposits of this glacier are separable from later glacial deposits on the basis of position and the greater depth of weathering. The oldest till lies in patches which cling to the valley walls, high above and downstream from the moraines deposited by the later ice. No terminal moraines of the early glacier have been preserved. Because of the character of this glacial debris and its topographic position, the writer believes that the Earlier epoch of Ball is correlative with the Wisconsin I substage.

The Wisconsin II substage is termed the Later glacial epoch by Ball (1908). Ice tongues of this substage were slightly smaller than those of the Wisconsin I substage. In Clear Creek Valley, the terminal moraine of the Wisconsin II substage forms a morainic complex at an elevation of approximately 8500 feet, about 3 miles up the valley from the terminus of the Wisconsin I glacier. At the junction of the valleys of West Fork and Clear Creek, near Empire mining camp and village, a morainic mass of till and associated stratified gravel was mapped by Ball (Spurr and Garrey, 1908, Pl. 80). According to his interpretation, the terminus of the Clear Creek glacier spread across the valley of the West Fork, forming a dam. Between the ice margin and the terminus of the West Fork glacier, a little more than a quarter of a mile upstream, a kamelike mass of stratified outwash was deposited. The complex hummocky topography at the mouth of Clear Creek suggests that the glaciers remained at this advanced position for a long time, with minor advances and retreats of only a few hundred feet.

In the valleys of Chicago and West Chicago creeks, tributary to Clear Creek, moraines of the Wisconsin III substage are located at elevations of approximately 8900 and 9200 feet respectively. Large lateral moraines in both valleys dam small unglaciated tributary valleys to form the famous Echo Lake on the east valley wall of Chicago Creek and Edith Lake on the west valley wall of West Chicago Creek.

Ball recognized that these lateral moraines are complex and intimated that they may represent more than a single ice advance. He says (Ball, 1908, p. 34),

> "These moraines are in some places double, the outer strand containing débris of more local origin."

Immediately below the Chicago Lakes, the lower morainic ridge, below the lateral moraine assigned to the Wisconsin II substage, merges with a low and rather poorly defined terminal moraine at an altitude of approximately 10,500 feet. A reservoir occupies the valley floor immediately above this terminal moraine, which is interpreted as belonging to the Wisconsin III substage, when a glacier almost 4 miles long moved down the valley.

The innermost and lowest lateral moraine on the west side of Chicago Creek Valley appears to merge into a poorly defined terminal moraine at an elevation of approximately 11,100 feet, recording the existence of a small ice tongue that pushed out from the cirque for about 2 miles. This small ice tongue is interpreted as the Wisconsin IV substage.

The cirques at the head of Chicago Creek were not entered by the writer. However, a heavy accumulation of talus near the cirque headwall above Upper Chicago Lake was seen from above. It did not appear to be a protalus rampart.

No attempt is made here to interpret the great bedrock step, about 400 feet high, which marks the lip of the eastern cirque containing Upper Chicago Lake. Similar giant bedrock steps are a feature of certain cirques irregularly located in the Colorado Front Range. Although they are apparently the result of glaciation, they appear to be related to local topography and rock type rather than to substages of glaciation.

In the valley of Clear Creek, at Silver Plume, at an elevation of 9000 feet, a mass of glacial debris reaches almost completely across the valley (Spurr and Garrey, 1908, Pl. 21). This is interpreted as a terminal moraine of the Wisconsin III substage and indicates that at a relatively late period in Wisconsin time there were a few comparatively large glaciers in these mountains. This particular glacier owes its length to the coalescence of a number of small ice tongues descending from the high cirques.

In the valley of the West Fork of Clear Creek, approximately 3 miles above Empire, at an elevation of about 8500 feet, there is a questionable moraine which may belong to the same Wisconsin III substage as the moraine at Silver Plume. If this is true, a glacier about 8 miles long occupied the valley of the West Fork. No attempt was made to explore the upper valley and cirques at the head of this creek. In the valley of Clear Creek, however, no striking morainal topography or debris was observed between Silver Plume and the Loveland Pass, with the exception of a small mass that is probably till at an elevation of approximately 10,750. If this can be interpreted as a moraine of the Wisconsin IV substage, it is reasonable to believe that similar moraines are located within the small tributary valleys of Clear Creek. In the cirques at Loveland Pass no protalus ramparts were observed.

In the Mount Evans region, there is definite evidence of four glacial substages of the Wisconsin. The Wisconsin V period of climatic refrigeration is not demonstrated. The massive character of the bedrock appears to be unfavorable for the formation of protalus ramparts. The unusual length of glaciers of the Wisconsin III substage in the valleys of Clear Creek and its West Fork is due to the greater area of accumulation and

the greater number of coalescing tributary glaciers than in regions farther north. The Chicago Creek Valley, whose area of accumulation is limited by the relatively small summit area of Mount Evans, had in this substage a very modest ice tongue.

PIKES PEAK

The great mass of Pikes Peak, standing apart and more or less isolated from the Colorado Front Range, reaches out into the Great Plains as a prong. Although Pikes Peak rises into the zone which may be expected to support small glaciers at the present time, its area of accumulation is so small, its position so unprotected, and the available rainfall so low that no permanent snowbanks lie on its summit. However, in Pleistocene time, the mountain supported several small glaciers.

Despite the fact that the summit of this peak has been accessible by tram and roadway for many decades, no attempt at a detailed study of the glacial geology has been made. Stone (1899, p. 348-349) writes that small glaciers once existed on the slopes of the peak. His information was doubtless derived from the brief account of glacial moraines given by Cross in the Pikes Peak Folio (1894). The latter recognized glacial moraines in the vicinity of Gillett, at the western foot of the mountain, where they are unusually well developed. Cross failed, however, to describe them in other parts of the area. Brief mention has been made of moraines on the eastern and northern side of the mountains (Matthes, 1910). Although the West Fork of Beaver Creek, above Gillett, was the only valley completely traversed by the writer, one can see evidences of the late glacial substages on the eastern and northern slopes of the mountain from the highway which leads to the summit.

On the eastern slope of Pikes Peak, Lake Moraine (elevation 10,215 feet) is held in by a terminal moraine (Finlay, 1916). This presumably marks the lowest elevation attained by glaciers on the eastern face of the mountain, where there are four large cirques. The automobile highway passes along the margin of one cirque headwall, known as the Bottomless Pit. Besides talus at the base of this cirque headwall, there are two small moraines, and on the northern valley wall a lateral moraine which is related to the lowest visible terminal moraine at an elevation of 10,500 feet (Fig. 6). This moraine is believed to represent the Wisconsin III substage. The youngest moraine, at an elevation of approximately 10,750 feet, is referred to the Wisconsin IV substage.

At least two moraines may be distinguished in Glen Cove, through which the automobile highway passes on its way to the summit (Fig. 6)—the lowest at an elevation of approximately 10,500 feet, the highest at an elevation of about 10,800 feet. The headwall of the Glen Cove cirque

contains a small mass of talus which seems to be a poorly defined protalus rampart. It is believed that the two moraines of Glen Cove represent the Wisconsin III and IV. The lower valley was not searched for older moraines.

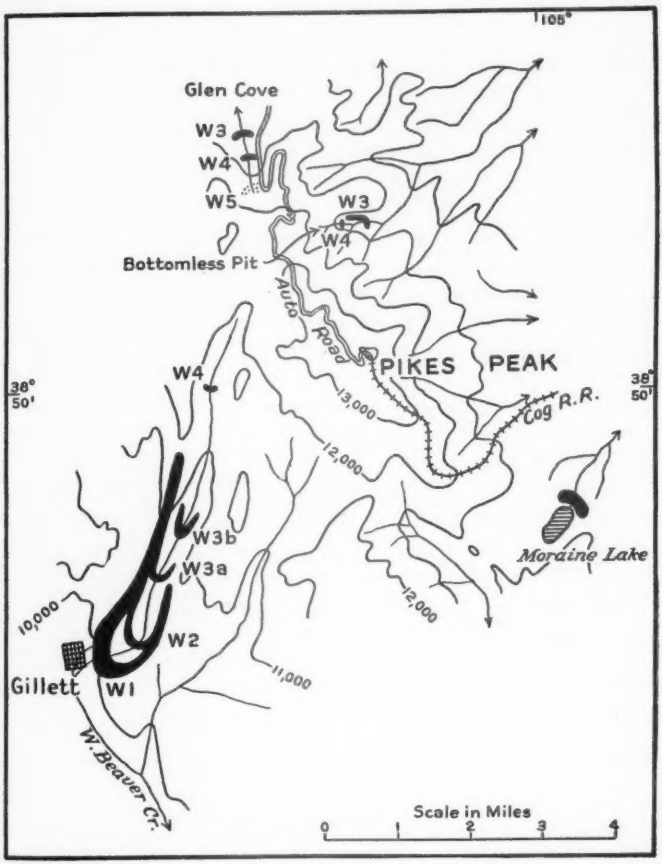

FIGURE 6.—*Sketch map showing position of the moraines observed in the Pikes Peak region*

In the valley of West Beaver Creek, east of the village of Gillett, a well-developed moraine at an elevation of 10,000 feet extends across the entire valley in a great loop. This moraine, recognized by Penrose as early as 1894 (Cross, 1894, special Cripple Creek map by R. A. F. Penrose), was

formed by a glacier approximately 6 miles long. It is believed to represent the Wisconsin I substage.

A great lateral moraine (Pl. 3, fig. 1), rising several hundred feet above the valley floor, is continuous with the Wisconsin I terminal moraine along the northern valley wall. Several smaller moraines lie on the lower flanks of this lateral moraine. These are continuations of later terminal moraines, which occur in the valley at elevations of approximately 200, 450, and 600 feet above the Wisconsin I moraine. Unfortunately, the till of these moraines is entirely composed of the coarse, reddish Pikes Peak granite, whose rapid weathering and inability to retain a glaciated surface make it difficult to distinguish degrees of weathering. It is also difficult to distinguish till from deeply weathered bedrock.

The writer believes that the Wisconsin II substage is represented by the moraine immediately upstream from and only 200 feet above the Wisconsin I moraine. Tentatively, the moraines at 450 and 600 feet above the Wisconsin I are assigned to the Wisconsin III substage. The Wisconsin IV substage is represented at an elevation of approximately 11,500 feet by a small mass of glacial debris perched on a bedrock "step" in the valley floor. Within the cirque there is much talus, but, as the cirque headwalls slope very gently and the cirque is broad, there is no indication of a protalus rampart.

It would seem, therefore, that during the Wisconsin stage the isolated Pikes Peak supported numerous small glaciers, which spread out in all directions from the high summit. At least eight glaciers moved down the mountain sides. One of the largest, in Beaver Creek on the western slope, was more than 6 miles long. Several substages of glaciation are indicated by well-defined moraines in this valley, which are interpreted as representing the first four Wisconsin substages. The Wisconsin V appears to be unrepresented except in the northward-facing Glen Cove cirque, where there is a small and poorly developed protalus rampart.

GLACIATION OF HUERFANO VALLEY AND WEST SPANISH PEAK

HUERFANO VALLEY

Although glaciation has long been recognized on Blanca Peak (Fig. 1), it has never been studied in detail. Siebenthal (1910), when working in the San Luis Valley to the west, made a reconnaissance of the mountainous country which culminates in the towering Blanca Peak (elevation 14,390 feet). He reported that many glaciers had formerly descended from the mountain crest, terminating far above the floor of the San Luis Valley. The longer valleys on the eastern side of Blanca Peak he believed contained larger glaciers than the shorter, more precipitous valleys on the

FIGURE 1. LATERAL MORAINE ALONG THE NORTH VALLEY WALL OF WEST BEAVER CREEK
Above the village of Gillett. This is primarily the product of the Wisconsin I substage.

FIGURE 2. VIEW FROM THE SURFACE OF THE WISCONSIN I MORAINE
Eastward down the valley of the Huerfano River. Note remnants of the old surface, above the present river valley, to which the Wisconsin I glaciers are related. Later substages of the Wisconsin are confined to the present river valley.

MORAINES

FIGURE 1. HUERFANO VALLEY
Showing lateral moraines of Wisconsin I and II substages. Photograph taken from high lateral moraine of the Wisconsin I substage. Lateral moraines of Wisconsin II substage occur on either side of valley in midground. Swampy meadow lies between Wisconsin I and II moraines.

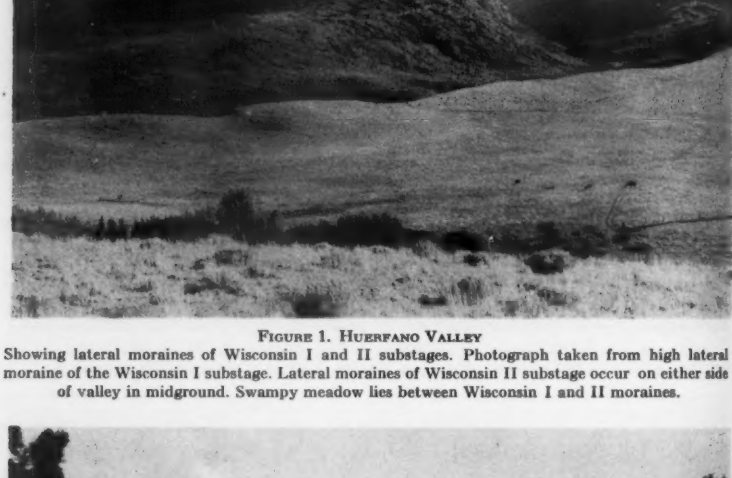

FIGURE 2. GLACIAL MORAINE OF THE WISCONSIN III SUBSTAGE IN WAHATOYA CANYON, WEST SPANISH PEAK
Coarse talus in foreground.

MORAINES

western slope. Siebenthal (1907) reported that a small glacier still existed at the head of the Huerfano Valley.

With these observations in mind, the writer examined the evidences of glaciation in the Huerfano Valley (Fig. 7). At an elevation of approximately 8800 feet (Huerfano Park Quadrangle), at the junction of the South and Middle forks with the main Huerfano River, a morainic complex of hummocky till is interpreted as belonging to the Wisconsin II substage, for above the river, on either side of the valley, lateral moraines rise to heights of 500 and 600 feet (Pl. 3, fig 2). These moraines rest on bedrock benches and are remnants of the more extensive and earlier glacial substage, the Wisconsin I. The old lateral moraines are hummocky on top, with numerous depressions inclosing small stagnant swamps. The morainal surface contains almost no large rocks. Most are deeply weathered, and occasional limestone fragments show solution facetting.

Downstream, high lateral moraines merge into terrace gravel that caps bedrock benches more than 150 feet above river level. There was a glacial advance before the completion of the "Canyon Cutting Cycle," during which a great ice mass occupied a position analogous to that of the Durango glaciation in the San Juan Mountains (Atwood and Mather, 1932). This glacier appears to have extended at least as much as a mile below the terminus of the Wisconsin II ice and to have terminated at an elevation about 200 feet higher. Farther up the valley of the Huerfano River (Fig. 7), where it swings to the south, the great Wisconsin I lateral moraine is separated from the Wisconsin II lateral moraine by more than 500 feet horizontally. Between these two moraines is a low swampy meadow of glacial outwash, almost 200 feet above river level. On the opposite side of the river are similar morainic systems (Pl. 4, fig. 1).

No evidence of wind-scoured pebbles or windblown sand and loess was found on the surface of the oldest moraine. The difference in topographic positions of the Wisconsin I and II moraines indicates the relative antiquity of the earliest Wisconsin substage. However, the degree of weathering and the topographic expression of the debris are not to be compared with the Cerro till of the San Juan Mountains or the Prairie Divide till of northern Colorado. This till is more akin to the Durango till at La Jara Reservoir in the San Juan Mountains, the earliest moraines in the Leadville region and the Medicine Bow Mountains, than to any of the pre-Wisconsin or Wisconsin II moraines and is therefore assigned to the Wisconsin I.

The fresher aspect of the till and the position of the moraines within the inner valley lead to a correlation with the moraines of the Wisconsin II substage. Deep cuts through this till show that weathering has penetrated little below the surface. Good fresh striated cobbles are common.

At an elevation of approximately 9100 feet, 1½ miles up the valley, is another well-developed terminal moraine of fresh till, which is assigned to the Wisconsin III substage. It is composed of uncompacted and unweathered white to light-gray till, with many striated cobbles and boulders.

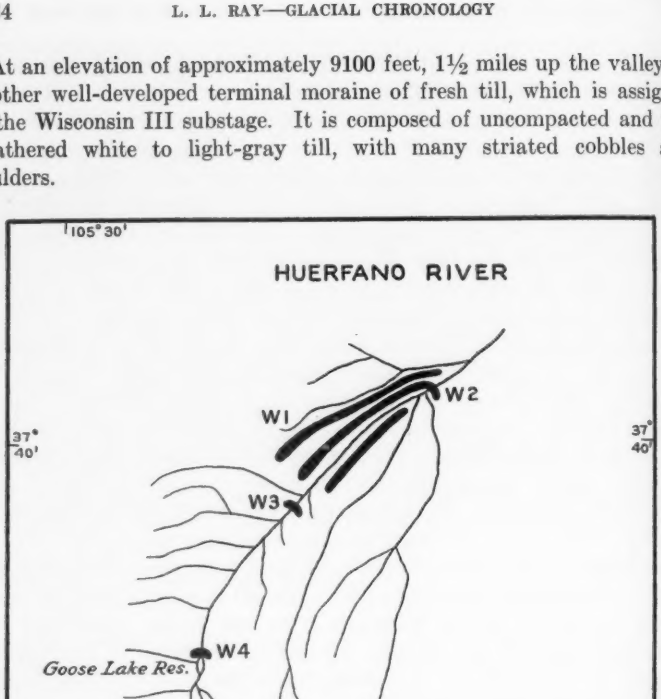

FIGURE 7.—*Sketch map showing position of moraines of the Huerfano Valley*

Between the Wisconsin III moraine and the now drained Goose Lake Reservoir (Fig. 7 and Huerfano Park Quadrangle) a mass of glacial debris appears on casual observation to be another terminal moraine at an elevation of 9600 feet. Close examination, however, shows that it can be interpreted only as a landslide, moving from the west valley wall into and partially across the main valley floor.

Not far upstream (Fig. 7), at an elevation of 10,200 feet, a glacial moraine interpreted as Wisconsin IV stretches across the valley. A small

dam, now washed away, was constructed across the valley at this moraine to form Goose Lake Reservoir, now a swampy meadow.

Although much glacial debris is strewn on the valley floor between the dam and the cirque lake, about 3 miles upstream, there is no definite moraine. Within the cirque, whose floor lies almost 500 feet above the main valley, there is a great accumulation of talus, whose hummocky form is similar to that of the rock glaciers of the San Juan Mountains, Evans Gulch near Leadville, Wheeler, and West Spanish peaks, and the near-by La Veta Pass. Adjoining cirques contain similar features. These talus accumulations appear to be in slow motion at the present time, indicating that the climate is still rigorous enough to furnish the required debris. If protalus ramparts were formed during the Wisconsin V substage, they have been obliterated by the rock glaciers, which have either buried or swept them away.

A permanent snowbank below the lip of the cirque and clinging under the precipitous crest of Blanca Peak has been called a glacier by Siebenthal (1907). It seems, however, that this mass of névé is not now a true glacier, for no glacial ice was found, and the meltwater is not silty glacial milk.

In the valley of the Huerfano River four definite substages of the Wisconsin glaciation are represented by moraines. A rock glacier was probably initiated during the rigorous climate of the fifth Wisconsin substage. The present climate is such that additions to the talus are sufficient to keep the rock glaciers in motion.

There is evidence in the valley of the Huerfano River of uplift and stream incision in the interstadial interval between the Wisconsin I and II substages. This is similar to the incision of streams in the San Juan Mountains, 50 miles to the west, and is the only area in the Southern Rocky Mountains where this marked canyon cutting during the Wisconsin stage was seen. Doubtless this region was affected by the same orogenic disturbance which was active in the San Juan Mountains.

WEST SPANISH PEAK

The two great peaks west of Trinidad, Colorado, known as East (elevation 12,708 feet) and West Spanish Peak (elevation 13,623 feet), are separated both topographically and geologically from the Sangre de Cristo Mountains immediately to the west. Hills' report (1901) on this region contains no statements relative to the Pleistocene geology. Glaciation was reported on West Spanish Peak by Salisbury (1901) and Knopf (1936). When viewed from the north, a cirque may be seen on the northeastern slope of this peak, at the head of Wahatoya Creek. The glacial features of this cirque and valley are here reported.

The cirque is easily reached by the Wahatoya Trail. The character of the lower slopes of the mountain and the lower valley are difficult to discern because of the heavy forest cover. Time did not permit examination of this obscured lower valley. It seems, however, on the basis of a forest-clad ridge stretching partially across the valley, that there is a terminal moraine at an elevation near 9000 feet.

The lowest till was seen at an elevation of 9800 feet, where good striated cobbles of typical soled and pentagonal form were collected. A morainic topography is poorly developed at this point. It is tentatively suggested that the oldest glacial substage, the Wisconsin I, reached an elevation near 9000 feet and that the Wisconsin II substage is represented by the till at an elevation of 9800 feet. More detailed study is necessary to prove the validity of these conclusions.

At an elevation of approximately 10,500 feet a great mass of till is perched on bedrock (Pl. 4, fig. 2). The till is relatively unweathered and preserves a distinct morainal form. It represents the Wisconsin III substage and marks the terminus of an ice tongue about three-quarters of a mile long.

The steep headwall of the cirque and the adjacent valley walls are cloaked by a thick talus mantel. Within the cirque is a large talus mass at an elevation slightly more than 11,000 feet (Pl. 5, fig. 1), whose outer ridge stretches across the cirque to form a barrier, which at a point near its western end has been cut by a stream channel. Here fine-grained glacial gravel and till are exposed. The presence of till suggests this is a moraine of the Wisconsin IV substage. During the substage when this till was deposited only a small ice mass could have existed within the cirque. However, the highly jointed bedrock facilitated the movement of debris from the cirque walls to the frontal margin of the ice.

Behind but close to this morainic ridge two ridges of talus parallel the valley walls almost at right angles to the barrier. These are interpreted as protalus ramparts, built in front of snowbanks which once encircled the base of the cirque walls. On the outer margin of these ramparts is a steplike series of ridges and benches which appear to result from the decreasing size of the permanent snowbanks.

Although the evidence for five Wisconsin substages in the Wahatoya Canyon on West Spanish Peak is not complete, it can be demonstrated that there were presumably that many. The earliest ice reached the lowest point in the valley at an elevation of approximately 9000 feet. The Wisconsin II substage reached an elevation of 9800 feet, the Wisconsin III 10,500 feet, and the Wisconsin IV and V were confined to the cirque, at an elevation of about 11,500 feet.

That this southern Colorado peak, so isolated from the mountains to

the west, should have supported glaciation is indicative of the cool, moist climate of the late Pleistocene. Furthermore, that it should also have recorded so many substages indicates that each of these was marked by a regional climatic shift, although the intensity of the change varied from substage to substage.

CULEBRA RANGE

Between the Spanish Peaks and the San Luis Valley to the west lies the Culebra Range of the Sangre de Cristo Mountains. Glacial phenomena in this region have been recognized and briefly mentioned by only two geologists—Siebenthal (1910, p. 38) and Upson (1938). Upson mentions the glaciation of the North Fork of Culebra Creek, pointing out that two great lateral moraines rise about 400 feet above the floor of the U-shaped valley. These he has called Wisconsin and tentatively correlated with the Wisconsin II here reported.

The northern portion of the eastern slope of the Culebra Range is accessible from Cuchara Camps and Stonewall and was the region chosen for examination. Numerous cirques scallop the eastern crest of the range, north of Trinchera Peak (elevation 13,546 feet). Since there are no topographic maps in this region, elevations are only approximate and were determined by aneroid barometer. A rough sketch map, based on the San Isabel National Forest map (United States Forest Service, 1937), has been constructed to show the general relationship of the glacial features (Figure 8).

Along the road leading from Cuchara Camps to Blue Lakes, the oldest morainic material occurs on the valley floor at an elevation of approximately 9900 feet, where the streams draining Blue Lakes join the South Fork of Cuchara Creek (Fig. 8). Here, good striated cobbles of red sandstone crop out in road side cuts through morainic ridges. Only small fragments of igneous and metamorphic rocks were found in this debris. They have been derived from Permian conglomerates in which the cirques have been cut and not from the underlying basal complex. This morainal material merges in a great lateral moraine (Fig. 8) which the writer believes represents the oldest glaciation of the Wisconsin stage. Unlike the Wisconsin I moraines in the Huerfano Valley to the north, it lies within the present valley. There is to be considered, of course, the possibility that some of this till may represent an ill-defined terminal moraine of the Wisconsin II substage and that the Wisconsin I lateral moraine lies on a bedrock bench, its relationship to bedrock obscured by the later morainic debris of the Wisconsin II substage which may mantle the lower slopes.

The Wisconsin II ice advance is marked by a small lateral moraine plastered on the side of the larger moraine (Fig. 8). This inner moraine,

however, is ill-defined, and the terminal position of the ice of this substage is unknown, unless it forms the lowest debris within the present valley, at an elevation of 9900 feet. The ice of this second advance did, however, descend at least far enough to cause the damming of the Bear

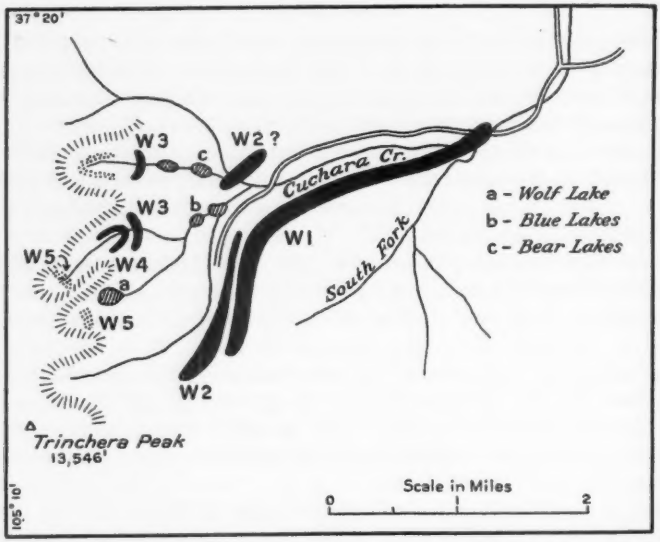

FIGURE 8.—*Sketch map showing positions of the glacial moraines at the head of Cuchara Creek, Culebra Range, Colorado*

Lakes by morainic debris (Fig. 8). The heavy timber of the lower valley prevented determination of the character and extent of the glacial deposits.

Below the lips of the three cirques north of Trinchera Peak are large masses of hummocky morainal material, shown diagrammatically in Figure 8. This jumbled mass of glacial debris is interpreted as material deposited during the Wisconsin III substage, when glaciers moved only a short distance from the mouths of the cirques. The debris is not of typical glacial form, and even faintly striated cobbles are rare.

Within the center cirque, where there is the best glacial sequence, a small ice tongue pushed to the cirque lip after the deposition of the scattered debris of the Wisconsin III substage. Shoving aside the older debris it left a looped ridge, interpreted as the product of the Wisconsin IV substage. Within the adjacent cirques to the north, there is no evi-

dence of such an ice advance, and it may be that in the later Wisconsin time these cirques were devoid of glacial ice.

Small protalus ramparts were developed after the final disappearance of the glaciers themselves in these cirques. Landsliding seems to have augmented the protalus material. Pl. 5, fig. 2 shows the third cirque north of Trinchera Peak, where a protalus rampart follows the main axis of the cirque along the northern valley wall. Along the southern valley wall only a minute rampart lies in front of the talus of recent origin.

It is apparent that the Culebra Range was glaciated during several substages of the Wisconsin. On the western slope a Wisconsin II moraine has been observed. On the eastern slope, glaciation representing five Wisconsin substages can be found, although not in each clique. Glaciers extended at least as low as 9900 feet and perhaps to still lower elevations, as yet undetermined.

GLACIATION IN NEW MEXICO

GENERAL STATEMENT

Although it has long been known that certain mountainous regions in New Mexico supported small glaciers during the Pleistocene period, little attempt at any systematic study of this glaciation has been made. As early as the time of Stevenson's report for the Wheeler Survey (1881) certain high summits of the Sangre de Cristo Mountains were recognized as having supported small glaciers. Salisbury (1901) and Stone (1901) reported briefly on glacial deposits in these mountains. Ellis has described the glaciation of the Moreno and Red River valleys (1931) and the glaciation of all New Mexico (1935).

Those mountain areas in New Mexico which were high enough to support glaciers during the Pleistocene may be divided into three main groups: (1) the Sangre de Cristo Mountains of north-central New Mexico (Fig. 9), (2) the San Juan Mountains and the Canjilon area (Atwood and Mather, 1932; Smith, 1937), and (3) the Sierra Blanca Mountains of south-central New Mexico. Antevs (1935, p. 306-307) has suggested the possibility of glaciation on this peak.

The glaciation of the Sangre de Cristo Mountains was studied in four different areas, which will be described in order: (1) the Moreno Valley, (2) Wheeler Peak, (3) Jicarita and Jicarilla peaks, and (4) Lake Peak (Fig. 9).

No systematic search was made to uncover in New Mexico evidence of a glacial stage older than Wisconsin. The writer believes, however, that a stage probably existed in the Sangre de Cristo Mountains similar to that in the mountains to the west of the Rio Grande Valley, as described by Ellis (1935), Smith (1937), and others. The material on

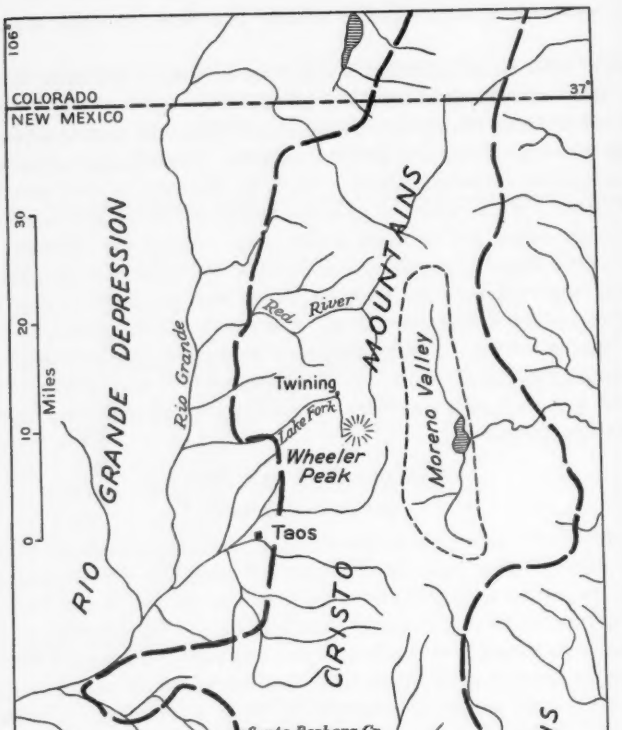

FIGURE 9.—*Index map showing location of regions studied in New Mexico*

U. S. Hill (State Highway 3, southeast of Taos), described by Ellis (1935, p. 14-15) as glacial till of Cerro age, was examined, and a test pit was dug. The deposit consists of rounded pebbles and cobbles of quartzite and granitoid rocks imbedded in a matrix of sticky clay, ranging in degree of weathering from slightly to highly decayed. Many of the pebbles are fractured, and the fragments slightly separated. No pebble or cobble could be found which had either glacial striations or a truly glacial shape. This deposit does not appear to be of glacial origin and is believed by the present writer to be a deeply weathered outcrop of the Picuris formation (Cabot, 1937). Areas of Cerro till shown by Ellis (1935, Pl. 4) at the head of Rio Fernandez de Taos, and near the village of Red River, are not till but seem to be either the weathered outcrop of a Tertiary formation—perhaps the Picuris—or a high-level gravel deposit related to the high pediments of the Rio Grande Valley. If till of Cerro age is found in the Sangre de Cristo Mountains, it will probably be on some of the high, flat surfaces within the mountains, at elevations of approximately 10,000 feet.

MORENO VALLEY

The broad, flat Moreno Valley (Fig. 9) lies within the rugged Sangre de Cristo Mountains and is roughly parallel to the main trend of the range. To the west lie high peaks such as Wheeler, Gold Hill, and Van Deist; to the east lie Baldy Peak and numerous other lower peaks.

Glaciation of the Moreno Valley has been described in detail by Ellis (1931; 1935), who believed that at one time it contained a large, southward-moving ice tongue fed from the high summits on either side and to the north. He postulated an ice lobe pushing from the western margin of the main Moreno Glacier across the divide between the Moreno and the Red River valleys. Ellis' map of 1935 shows this glacier occupying the Moreno Valley and extending southward possibly beyond Black Lake, whose elevation is 8526 feet—several hundred feet above the lowest point in the Moreno Valley to the north, at the mouth of the Cimarron Canyon. The lowest point at which glacial ice has ever been reported in New Mexico is 8500 feet.

The ice lobe which pushed into the Red River Valley as shown by Ellis (1931) extended as far downstream as the village of Red River. Here, however, the glacial debris is not supposed to occur on the valley floor but to lie 500 to 700 feet above it, on rock terraces or berms. This is confusing, for the map of 1931 indicates that these deposits belong to the same glaciation as that of the Moreno Valley, but the map of 1935 shows them to be Cerro in age, although the main glacier in the Moreno Valley is indicated as Wisconsin. As Ellis states that the Red

River has cut a gorge several hundred feet below the glacial deposits (Ellis, 1931, p. 14-25), there is little doubt that he changed his beliefs between 1931 and 1935. In 1931 he stated that this terrace probably "marks the beginning of ice retreat" (Ellis, 1931, p. 24). On the map accompanying his paper of 1935 these deposits are correlated with the Cerro glaciation, thus indicating that the stream incision was accomplished after the Cerro glaciation and not after the Wisconsin glaciation.

The present writer spent several days in the Moreno and Red River valleys in 1938 and in 1939 he spent an entire field season in the Moreno Valley. Detailed study shows that the Moreno Valley, below Red River Pass, and the Red River Pass itself were never occupied by a glacier in Wisconsin time, and no evidence was seen which would indicate that it had ever been occupied by a glacier.

The Moreno Valley has no ground moraine scattered over its broad floor. Where road cuts have been made along the valley floor only gravel deposits that are related to the terraces are exposed. Ellis (1935, p. 8) reports,

> "what seems to be a series of recessional moraines extending nearly across the valley, from east to west. . . . The material of these ridges consists of gravel and boulders of granite, gneiss, quartzite, etc., with a predominance of local rocks. In some places the material shows a slight stratification, and in some respects the ridges resemble an eroded terrace. On the other hand, the condition of the intervening valleys (they are separated by marshy land) is more consistent with glacial action."

These "recessional moraines" can be interpreted only as terrace remnants.

Along the eastern valley wall, immediately north of the dam which forms Eagle Nest Lake (Fig. 9), there is a ridge of hummocky form which Ellis (1931, p. 26) interpreted as glacial in origin. This ridge, however, has been carefully examined and is not composed of glacial debris but of deformed Tertiary deposits, shale, sandstone, igneous intrusives, and pre-Cambrian gneisses. From a distance, the grassy slopes and hummocky topography give an impression of morainal topography which may be mistaken by the incautious observer.

Near the village of Red River, no signs of any glacial deposits were seen. Although no Wisconsin glacier extended as low as 8600 feet in the present inner valley, it is not certain that a small pre-Wisconsin glacier might not have deposited the debris described by Ellis on the remnants of the old berms. On the other hand, it is reasonable to believe that the gravel—no till is mentioned by Ellis—has been deposited during the period of formation of the high pediments in the Rio Grande Valley.

As the writer has found no evidence of either a Red River lobe or a main valley glacier moving southward in the Moreno Valley, he is con-

FIGURE 1. LARGE TALUS MASS WITHIN CIRQUE AT HEAD OF WAHATOYA CANYON, WEST SPANISH PEAK
A small stream channel cut into this mass reveals a core of glacial till.

FIGURE 2. CIRQUE ON EASTERN SLOPE OF CULEBRA RANGE, NORTH OF TRINCHERA PEAK
Showing protalus ramparts of the Wisconsin V substage. (*See* Bear Lakes cirque, Figure 8.)

FIGURE 3. SMALL CIRQUE ON THE EAST VALLEY WALL OF THE SOUTH FORK OF LAKE CREEK
Below cirque lip is debris of the Wisconsin IV glacial substage. Within cirque is small protalus rampart which represents the Wisconsin V substage.

FIGURE 1. CIRQUES ON THE EASTERN SLOPE OF THE SANGRE DE CRISTO MOUNTAINS NORTH OF WHEELER PEAK

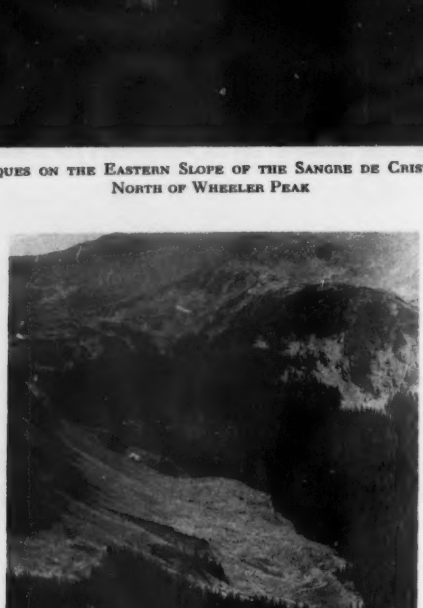

FIGURE 2. TALUS MATERIAL
Assuming forms characteristic of rock glaciers along the southern valley wall and immediately below the lip of Wheeler Peak cirque.

vinced that, although the high peaks of the Sangre de Cristo Mountains supported small glaciers in late Pleistocene time, the glaciers were not large enough to descend from the high mountains to coalesce into a single ice tongue large enough to move down the broad Moreno Valley.

WHEELER PEAK REGION

General statement.—Wheeler Peak (elevation 13,150 feet), 13.5 miles northeast of Taos, is the highest summit of the northern portion of the Sangre de Cristo Mountains in New Mexico (Fig. 9). Because of its height and the accessibility of the cirques on its flanks, two of the major stream valleys, one heading on the east and the other on the west side of this peak, were explored for evidence of glacial substages. The valleys of the East Fork of Red River and the Lake Fork of the Rio Hondo were followed to their sources.

Lake Fork of Rio Hondo.—An automobile road follows the canyon of the Rio Hondo to Twining, a small mining camp at an elevation of 9400 feet (Fig. 10). There are no evidences of glaciation to this point in the canyon. At several localities, however, heterogeneous, angular debris crops out along the valley bottom. Casual observation might lead to the belief that it is of glacial origin, but closer inspection indicates that it is landslide.

Above Twining (Fig. 10), morainic debris was first seen at an elevation slightly below 10,000 feet. Continuous hummocky topography and glacial till extend to an elevation slightly above 10,500 feet, where there is some indication of a morainic barrier across the valley. This material may represent either a single substage of the Wisconsin glaciation or two distinct substages. The writer is inclined to believe that the debris represents two distinct glacial substages—the Wisconsin I and II. However, no recognizable differences in the depth of weathering of these tills could be detected. This portion of the valley is clothed with a heavy stand of timber so that it is impossible to gain a clear picture of the relationship between the tills. There is, however, no mistaking the till, for good striated cobbles were found, and here and there a bedrock outcrop showed a roche moutonnée surface.

On the cirque lip, at an elevation of 11,000 feet, a great morainic dam stretches across the valley. It is composed of scattered piles of angular blocks which hold in the waters of Williams Lake but through which water seeps to form the source of Lake Fork (Fig. 10). The possibility that this dam is a landslide was considered in the field but rejected because the valley is too wide for the development of such an accumulation. Lack of striated cobbles on the surface of this moraine can be accounted for by the short distance of transportation from the cirque

headwalls and that the greater portion of this debris was probably carried on the surface of the ice.

The cirque at the head of Lake Fork was the largest seen in New Mexico. The easily fractured bedrock readily permitted plucking by

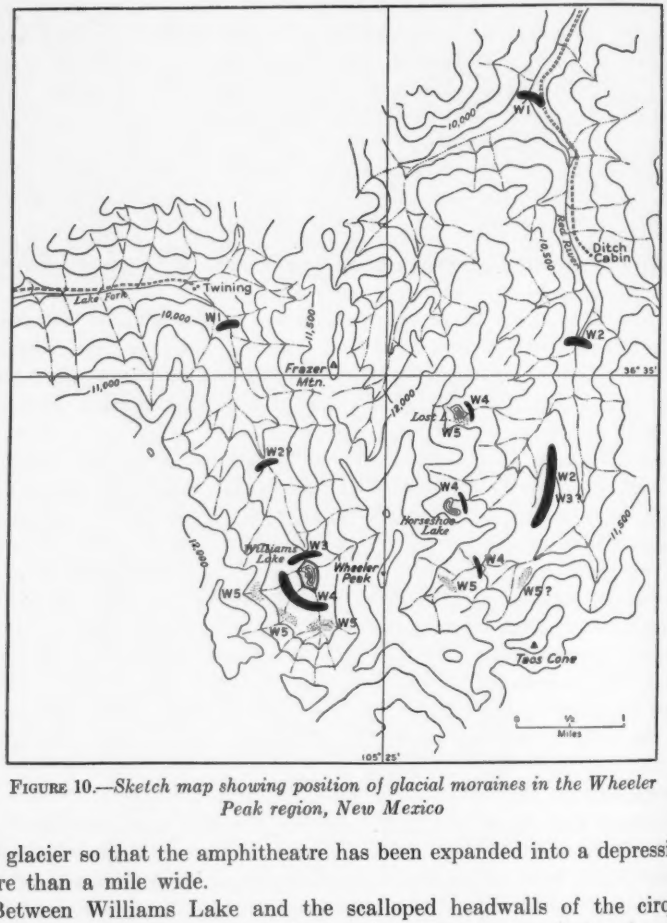

FIGURE 10.—*Sketch map showing position of glacial moraines in the Wheeler Peak region, New Mexico*

the glacier so that the amphitheatre has been expanded into a depression more than a mile wide.

Between Williams Lake and the scalloped headwalls of the cirque many irregular masses of morainal material are the products of small "glacierets" which were the last ice remnants. The cirque headwalls are talus covered, and in places there are small protalus ramparts.

Morainic debris at the mouth of the cirque is assigned to the Wisconsin

III, the small moraines within the cirque, shown diagrammatically in Figure 10, to the Wisconsin IV, and the protalus to the Wisconsin V, the last period of climatic refrigeration. Today small snow masses rest in sheltered nooks of the southern cirque wall, where they are protected from the sun and remain until late summer. It is not improbable that after winters of unusually great snowfall some of the snow may remain throughout the summer.

In the valley of Lake Fork are evidences of five substages of the last glaciation. The first or Wisconsin I is represented by material deposited at an elevation of about 10,000 feet by a glacier approximately 3.25 miles long; the Wisconsin II was deposited at an elevation of about 10,500 feet by a glacier slightly more than 2 miles long. The Wisconsin III, at an elevation of 11,000 feet, was deposited by an ice mass too small to push from the cirque. The last two feeble accumulations are within the cirque—the Wisconsin IV, the accumulations of glacierets which occupied the scalloped border of the cirque headwall, and the Wisconsin V, the result of nivation during the last period of climatic refrigeration, which was too weak to bring about the formation of glacial ice capable of movement.

Red River Valley.—The East Fork of Red River drains the cirques on the eastern side of Wheeler Peak and the long ridge which runs northward from the main mountain mass (Fig. 10). Throughout the lower valley, from Red River Pass to the junction of the East and West forks of Red River, numerous well-developed terrace remnants are remnants of valley trains, correlative with glacial advances. At the junction of the forks, at an elevation of approximately 9400 feet, there is hummocky morainal debris in the bottom of the valley. Here is the incompletely preserved and lowest moraine, the Wisconsin I, deposited by an ice tongue about 6 miles long.

About three-quarters of a mile above the Ditch Cabin at an elevation of 10,000 feet (Fig. 10), there are exposures of glacial till along the stream containing numerous striated boulders, with typical pentagonal and soled glacial form. This deposit is the Wisconsin II terminal moraine.

Above the Wisconsin II moraines a small post-glacial box canyon is cut in the limestone bedrock of the valley floor to a depth of about 12 feet. When climbing from the main valley to the Lost Lake cirque (Fig. 10), through dense forest growth, one is unable to recognize any distinct glacial features. However, on viewing the valley from the small moraine perched on the cirque lip, a great forest-covered ridge is visible which may be interpreted as a lateral moraine of either the Wisconsin

I, II, or III substages. It is the only glacial feature found which might possibly be assigned to the Wisconsin III.

Glacial debris on the lip of Lost Lake cirque, at an elevation of 11,300 feet, is interpreted as Wisconsin IV. Lost Lake is only partially held in by this debris and occupies a basin in the limestone and underlying pre-Cambrian igneous rock of the mountain core. Along the southern cirque wall, in the area most shaded and protected from the sun, are talus slopes formed largely of blocks derived from the sedimentary rocks of the cliffs above. A vegetation-covered protalus rampart lies in front of the talus slope of the southern cirque wall. Talus is now filling in the fosse between the protalus rampart and the cirque wall. The western cirque wall, which is relatively unprotected from the sun, is covered with a thin talus layer. At present small channels are being cut into the talus, indicating that the equilibrium recently attained is now being upset.

Two smaller cirques lie between the Lost Lake cirque and the great cirque on Wheeler Peak to the south. One of these contains Horseshoe Lake (Fig. 10; Pl. 6, fig. 1). None of these were entered but were viewed from across the valley. From a distance small ridges within the cirques appear to be correlative with the debris on the lip of the Lost Lake cirque or the Wisconsin IV substage. Scattered debris in the lower valley may belong to the Wisconsin III substage; the moraines of the Wisconsin II and I substages lie farther down the valley at 10,000 and 9400 feet, as previously described.

The headwall of the Wheeler Peak cirque is not precipitous but slopes gently. Debris is probably now moving slowly down the gentle slope behind the protalus rampart and has assumed the form of great garlands. Below the protalus rampart a small moraine perched on the low rock lip of the cirque is correlated with the Wisconsin IV, the protalus rampart with the Wisconsin V. Outside the cirque proper but along the southern valley wall festooned talus accumulations (Pl. 6, fig. 2) are similar to the rock glaciers described by Howe (1909) from the San Juan Mountains and by Patton (1910) from the La Veta Peak region to the north. In form and location these masses are similar to those of the Lost Lake cirque but lack the definite vegetation-covered protalus rampart. Without doubt the rock glaciers in the Wheeler Peak region have resulted from the oversteepening of the valley walls by glaciation and from the intense frost action during the last or Wisconsin V period of climatic refrigeration. At present they apparently are receiving additional talus in sufficient amounts to keep them in motion so that vegetation has not yet gained a foothold on their surfaces.

The eastern slope of Wheeler Peak holds a glacial sequence similar to that found on the western slopes. On the eastern side the Wisconsin I,

II, IV, and V are well developed and easily distinguishable. The Wisconsin III has been tentatively defined on the basis of the topography of the heavily forested valley floor. The existence of the Wisconsin I rests on the character of the debris of the valley floor near the junction of the East and West forks of Red River. On the basis of the position of this debris, the maximum length of the glacier descending from the great cirque on the east side of Wheeler Peak would be about 6 miles, almost twice the length of the glacier in the valley of Lake Fork. The floors of the cirques on the east side of the peak are several hundred feet higher in elevation than those on the west side, which does not seem in harmony with the differences in the length of the glaciers. However, this phenomenon may be due to differences in bedrock.

JICARITA AND JICARILLA PEAKS

Southeast of Taos, near the headwaters of Santa Barbara Creek (Fig. 9), lies a group of high peaks known by the general name of Truchas Peaks. The northernmost of the Truchas Peaks is Jicarilla (elevation 12,944 feet). About 5 miles northeast of Jicarilla Peak is Jicarita Peak (elevation 13,100+ feet). Truchas Peaks and Jicarita are connected by a high ridge, which in this region forms the backbone of the Sangre de Cristo Mountains. Both the Truchas Peaks and Jicarita have been cited by Ellis (1935, p. 16) as having been glaciated, but neither was closely examined by him.

On the East Fork of Santa Barbara Creek, the lowest moraine observed lies slightly above an elevation of 9500 feet, immediately above the junction of the East and West forks (Fig. 11). Two small cirques on the west valley wall of the Middle Fork, just above its junction with the East Fork, have small moraines perched on their lips. These cirques are not distinguishable on the topographic maps because of their small size. The lowest moraine in the valley, just above 9500 feet, is interpreted as Wisconsin II. The Wisconsin III is represented by a moraine at an elevation of 10,000 feet. Heavy vegetation prevented determination of the shape of the glacial debris, which was seen only in exposures along the stream.

At an elevation of 10,500 feet, glacial till, interpreted as the Wisconsin IV moraine, occupies the valley floor. Dense forest prevented an areal examination of its extent and topographic expression.

The cirque floor was not examined. It is well forested and seems to contain no definite ridge which might be termed a moraine. Where visible, the headwalls are partially masked by talus, perhaps the product of the Wisconsin V substage but more likely of recent origin. Airplane photographs reveal no feature in the cirque which might be considered a moraine. They do show, however, that below the junction of the East

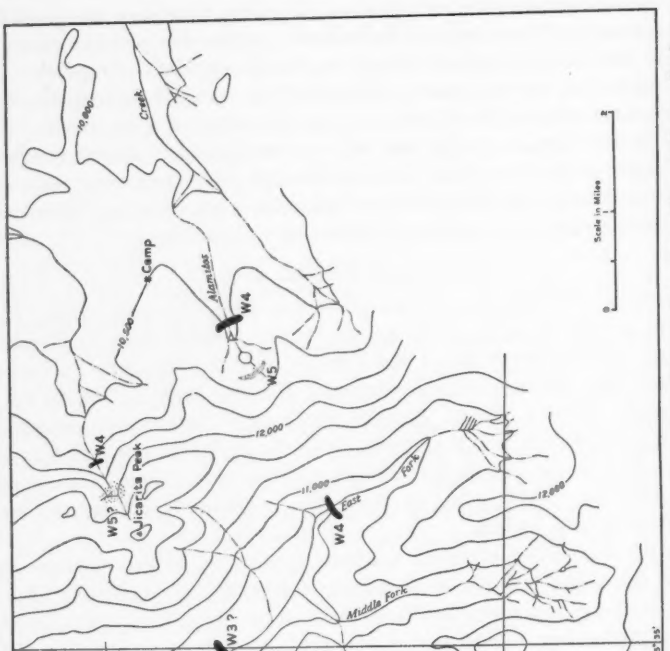

FIGURE 11.—*Sketch map showing position of the moraines observed in the valley of Santa Barbara Creek, New Mexico*

and West forks the Santa Barbara Valley changes from a broad U-shape to a narrow V-shaped gorge. This change is due to the combination of glaciation and structural control.

On the West Fork of Santa Barbara Creek (Fig. 11), a terminal moraine occurs at an elevation of 9600 feet, slightly over a mile above the junction of the forks. This is believed to represent the Wisconsin II substage, the correlative of the moraine seen at 9500 feet on the East Fork. Along the east side of the valley, above the moraine, is a small lateral moraine whose top lies about 125 feet above the present valley bottom.

In a reconnaissance trip up the West Fork of Santa Barbara Creek, no terminal moraine was seen which would be comparable to that seen on the East Fork at an elevation of approximately 10,000 feet. It is assumed that in the West Fork the Wisconsin III moraine was either not developed or is so obscured and modified that it is not recognizable. Nothing which could be termed a moraine was seen between the Wisconsin II moraine at 9600 feet and the great rock "step" at the mouth of the cirque.

Perched on the cirque lip and extending more or less across its roche moutonnée surface is a body of glacial debris which is interpreted as the moraine developed during the Wisconsin IV substage. This material is about three-quarters of a mile below the headwall of the cirque.

Unfortunately, a storm broke when the lip of the cirque was reached, and low hanging clouds prevented further exploration of the cirque itself. No moraine could be seen within the cirque. However, the northern exposure of the cirque may have been favorable to the accumulation of snow, resulting in the formation of protalus ramparts during the Wisconsin V substage.

Four large cirques nestle on the east flank of Jicarita Peak and on the ridge extending to the south (Fig. 11). These were not examined in detail but were observed from the high ridge separating the two northernmost cirques. No attempt was made to explore the lower courses of the streams draining from these cirques to determine the lowest position of the moraines. From distant views, it appears that the lower valleys were glaciated to altitudes similar to those attained by the glaciers of the Santa Barbara Creek region. In these cirques there is much talus of limestone and in the northernmost cirques low ridges covered with shrubs indicate protalus ramparts in front of which there are low moraines interpreted as belonging to the Wisconsin IV substage. In general the floors of the cirques lie at elevations of about 10,500 feet, slightly lower than the floors of the cirques on the western side of the mountains.

Two small cirques on the northwestern side of Jicarita Peak were noted by Ellis (1935, p. 16). They were seen by the writer only from the road near the village of Llano, but careful examination would doubt-

lessly reveal that they displayed a glacial sequence similar to that of the near-by cirques.

In the region of Jicarita and Jicarilla peaks there is evidence for several substages of glaciation, similar to those of the regions farther north, in the Colorado Front Range. Substages referred to the Wisconsin II, III, and IV are best represented. No evidence of the Wisconsin I was found in the Santa Barbara Creek valley. It has presumably been removed by erosion during the Wisconsin I-Wisconsin II interstadial period or by the advance of the Wisconsin II ice. Large quantities of talus in the cirques indicate that in favorable locations the last period of climatic refrigeration was such that small protalus ramparts were built. These represent the Wisconsin V substage.

LAKE PEAK

Lake Peak (elevation 12,380 feet), the southernmost peak of the Sangre de Cristo Mountains to be examined for evidences of glaciation, lies about 12 miles northeast of Santa Fe (Fig. 9).

Santa Fe Creek drains from a large cirque on the southeastern side of Lake Peak. This cirque contains a small lake (*see* Santa Fe Quadrangle), held in by a low morainic ridge. The lake, according to the topographic map, has an elevation of about 11,500 feet. About 200 feet below and a quarter of a mile down the valley from this first morainic dam is another small moraine seen only from a distance. Within the cirque there is no evidence of a protalus rampart, and the writer believes that during the last period of climatic refrigeration, when such ramparts were being formed in the mountains to the north, the open southward exposure of this cirque and the massive character of the granitoid bedrock were unfavorable for the accumulation of permanent snowbanks and the building of a protalus rampart from material derived from the cirque headwall. The lower valley of Santa Fe Creek was not examined, and elevations of lower moraines not determined, but Salisbury (1909, p. 729) reports a moraine at an elevation of 9200 feet.

On the northwest slope of Lake Peak is a small cirque whose floor lies at approximately 11,000 feet. From this now-forested cirque a small glacier descended to an elevation of approximately 10,000 feet, where it built a moraine. The coarse-grained, granitoid rocks on the surface of this moraine have failed to preserve their glaciated character and show no signs of striations. The moraine, however, preserves its hummocky form and stretches partially across the valley. A smaller moraine occurs farther up the valley at an elevation of about 10,300 feet. It is not so well developed as the older moraine at 10,000 feet and does not completely block the valley. At the lip of the small cirque and immediately

below is a formless mass of detritus accumulated by the last ice mass to occupy the cirque. Within the cirque there is a little forest-covered talus, but there is no suggestion of a protalus rampart.

There are on Lake Peak evidences of at least three glacial substages, which are here interpreted as Wisconsin II, III, and IV. A more detailed study might reveal remnants of an older glaciation, the Wisconsin I. The Wisconsin V was never developed.

Numerous small cirques may be seen in the higher parts of the mountains near Lake Peak but they were not examined. From the village of Cebolla (Las Vegas Quadrangle), several small cirques were seen which nestle below the crest of the eastward-facing escarpment of the more or less flat central mountain highland. On the map of this region the contour lines are so generalized that these cirques do not show, nor is the glaciated character of the country apparent. The cirque floors along the Cebolla escarpment seem to be at elevations slightly less than 11,500 feet. They were doubtlessly fed by snow blown across the great upland flat to the west.

LEADVILLE—ASPEN REGION

UPPER ARKANSAS VALLEY

As early as 1869, J. D. Whitney of Harvard University led a field party of students, among who was W. M. Davis, into the Sawatch Mountains, just south of the region here considered. In 1873 and 1874 the Hayden Survey made a geological reconnaissance through the Upper Arkansas Valley. Mining development in Leadville and Aspen has led to numerous studies of the economic geology. Usually these reports have included a brief chapter on the topography and surficial deposits.

The publication of the Leadville Reconnaissance Map in 1891 emphasized the highly glaciated character of the mountains, which had been recognized by the earliest geological surveys. The magnitude of the glacial phenomena, shown in the high mountains and expressed in the great terminal moraines in the Arkansas Valley, led to a detailed examination of the Pleistocene geology of the Leadville region by Capps (1909) and studies of such areas as that about the famous Twin Lakes, at the mouth of Lake Creek Canyon, by Davis (1905) and Westgate (1905).

More recently, the Upper Arkansas Valley has been studied by Behre (1933b) and Powers (1935), whose papers have dealt primarily with the topographic development of the region. The writer here gives a reinterpretation of the previous studies based on a brief examination of this region in September 1938. Inclement weather and the short time available permitted visits to only the most accessible of the glacial valleys. However, on the basis of the country examined, it appears that the com-

mon sequence of glacial stages in the Front Range is duplicated in both the Sawatch Mountains west of the Arkansas Valley and in the Mosquito (Park) Range to the east.

Early studies in this region pointed out a single period of Pleistocene glaciation; each successive study has revealed a more complex glacial history. The early Hayden Survey found evidence for a single period of ice advance, which formed the great moraines of Clear Creek, Lake Creek, Half Moon Creek, and others. Later, Westgate (1905) and Capps (1909) recognized at least two periods of late Pleistocene glaciation, and both suggested the possibility of a third, older stage not confined to the present valleys. Behre, in his recent paper (1933), has suggested the possibility of another glacial substage, a period of formation of "recessional" moraines, which are found in several of the valleys. To this sequence the present writer adds two more recent substages—the Wisconsin IV and V.

The great terminal moraines spreading from the mouths of the canyons along the Arkansas River have been mapped by Capps (1909, Pl. 1). At several localities he found older drift immediately in front of these large moraines, as at Twin Lakes, near the mouth of Lake Creek. The youngest of these morainic systems has in the past been referred to the Wisconsin; the older has either been called pre-Wisconsin or simply older drift. The topographic position of these tills indicates, however, that they have been separated in time of deposition by a long interstadial period, which has not, however, been long enough to obliterate completely the morainal character of the surface of the older drift or to weather the materials as deeply as those tills assigned to the Cerro or Prairie Divide glacial stage. For this reason the writer assigns these tills to the Wisconsin stage, calling them Wisconsin I and II. The first Wisconsin substage called Twin Lakes derives its name from the oldest till at the mouth of Lake Creek (Capps, 1909, Pl. 1). The prominent moraines encircling the Twin Lakes are assigned to the Wisconsin II substage. At this point the Wisconsin I and II substages are in contact with one another (Fig. 12).

These tills are genetically related to terraces of the Arkansas Valley, and Behre (1933b, p. 798, 802) assigns them to his No. 2 and 3 terraces, which lie approximately 220 and 30 feet above the river. His highest or No. 1 terrace is not definitely correlated with a glacial stage, and he believes it doubtful if a glacial stage existed which could have produced it.

To the south, Powers (1935, p. 196-197) has described three moraines in the valley of Chalk Creek, near Mt. Princeton, which are also related to the terrace sequence in the Arkansas Valley. He has assigned these

moraines to the Wisconsin, the first pre-Wisconsin, and the "earliest" glacial stage and has correlated them with the 200-, 80 to 100-, and 60-foot terraces. It would seem to the present writer that all of Powers' moraines might be assigned to the Wisconsin Stage and may be termed Wiscon-

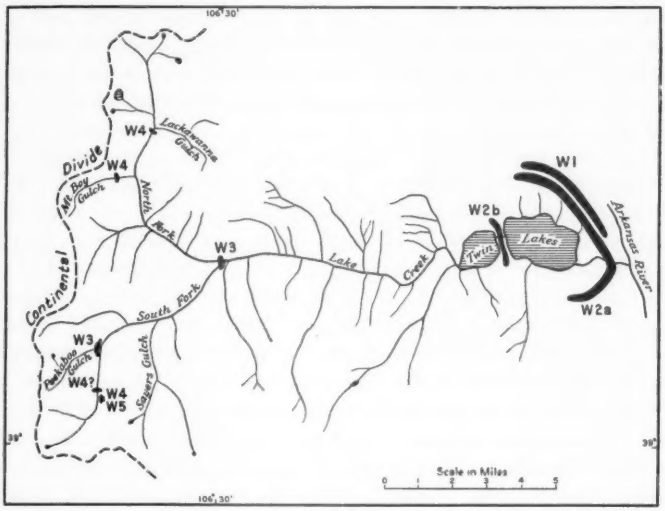

FIGURE 12.—*Sketch map showing position of the moraines of Lake Creek* South of Leadville, Colorado.

sin I, II, and III. The Wisconsin I and II are correlatives of the two morainic systems at the mouth of Lake Creek, south of Leadville. The oldest moraines, Wisconsin I, are believed to be correlative with the Durango glaciation of the San Juan Mountains, for they occupy analogous topographic positions and appear to have undergone approximately the same degree of weathering.

Behre (1933b, p. 790) notes that after the completion of the great terminal moraines, here called Wisconsin II, there are "distinct recessional moraines which appear in most of the valleys." Three examples are given (Behre, 1933b, p. 790): (1) three miles west of the mouth of Clear Creek, (2) at the Vega Shaft in Evans Gulch, at an elevation between 11,300 and 11,400 feet, and (3) in the valley of Homestake Creek, west of Pando. In Evans Gulch, the only one of these localities visited by the writer, there is a morainic complex about 600 feet lower in elevation and approximately 1½ miles down the valley from the moraine mentioned by Behre. It is to this lower moraine (Pl. 1, fig. 3)

that the Wisconsin III age is here assigned. The higher and younger moraine mentioned by Behre is held to belong to a more recent ice advance, the Wisconsin IV.

In the valley of Lake Creek, above Twin Lakes (Fig. 12), remnants of this Wisconsin III substage are in the form of low morainic topography and a small lateral moraine at the junction of the North and South forks. Ice of this substage descended to an elevation of about 10,000 feet. During its retreat in the South Fork of Lake Creek much debris was spread in hummocky masses throughout the valley, from an elevation of about 10,000 to 10,600 feet, or just beyond the mouth of Sayer's Gulch, a distance of about 3 miles. Similar topography was not developed in the valley of the North Fork. A lateral moraine formed by this ice tongue stretches across the mouth of Peekaboo Gulch, on the South Fork, at an elevation of 11,100 feet. Behind this lateral moraine lies a morainic complex thought to belong to this same Wisconsin III substage.

To the north, in the valley of Lake Fork Creek, there is no readily distinguishable Wisconsin III moraine. However, it may be that this stage is represented by the low hummocky deposit at an elevation slightly below 10,000 feet, described by Capps (1909, p. 43). Above it there is no evidence of any later stage of glaciation in the South Fork of Lake Fork Creek. The cirques at Bush (*see* Leadville Quadrangle) lack a heavy debris mantle and have no remnants of small moraines such as have been found in other valleys.

The youngest moraines found in the Leadville region are assigned to the Wisconsin IV substage of glaciation. In Evans Gulch such a substage is represented by the small moraine at an elevation slightly below 11,400 feet. This moraine is the one described by Behre (1933b, p. 790) as a later recessional stage. Here the low debris is a combination lateral-terminal moraine, poorly defined. Evidences of this substage were not seen in the valley of Lake Fork Creek but are developed in the valley of Lake Creek. At Lackawanna Gulch (Fig. 12) on the North Fork of Lake Creek a low moraine at an elevation of approximately 11,000 feet may be assigned to this Wisconsin IV substage. A correlative moraine occurs at the mouth of Mountain Boy Gulch, 2 miles downstream. Here, however, the moraine is perched at the mouth of the hanging valley and does not occupy a position analogous to that at the mouth of Lackawanna Gulch.

In the South Fork of Lake Creek, at an elevation of approximately 11,400 feet, there is a low hummocky mass of glacial debris which may be interpreted as a Wisconsin IV moraine. The valley above this moraine is broad and swampy, seemingly a remnant of an old filled lake. Below the moraine the stream has cut a small trench in the bedrock valley

floor, which is swept fairly clean of debris. There is some question as to whether this so-called moraine is not merely a landslide from the valley walls. This was considered in the field, and the landslide origin was rejected. No scars were seen on the valley walls from which this material would have been derived.

A small cirque on the east valley wall of the South Fork of Lake Creek (Pl. 5, fig. 3) has a mass of glacial debris just below its lip. This debris appears to be a small moraine deposited during the Wisconsin IV substage. Within the cirque is a protalus rampart developed during the Wisconsin V substage.

Similar masses of talus occur in many of the cirques of the Leadville region. In the Evans Gulch cirque is a well-known rock glacier whose origin may be traced to the rigorous Wisconsin V climate.

In the Leadville region, five distinct substages of the Wisconsin glaciation have been observed. On each major creek moraines of the Wisconsin II substage are present, and on some moraines of the Wisconsin I substage are preserved. These two morainic systems were formed by glaciers of approximately the same magnitude but were separated by a long interstadial interval. The older has been correlated with the Durango glaciation of the San Juan Mountains, and the younger with the "Wisconsin" of Atwood and Mather (1932). Two distinct younger substages have been recognized in Lake Creek Valley and Evans Gulch. These are correlated with the Wisconsin III and IV of the Front Range. Numerous cirques in the Leadville region contain debris which may be referred either wholly or in part to the Wisconsin V substage—a time of climatic refrigeration which was not great enough to produce definite ice tongues.

ASPEN REGION

In the mountainous regions surrounding Aspen, Colorado, across the Sawatch Mountains from the Arkansas Valley, extensive glaciers existed in Wisconsin time. Evidence of glaciation is abundant along the western slope of the mountains, where they are scalloped with well-developed cirques, many containing tarns. The valleys are strewn with drift, lacustrine deposits, and here and there terminal and lateral moraines. Roche moutonnée surfaces are common.

No published descriptions of glaciation in this region have been found, with the exception of the brief discussion by Spurr (1898). He described a glacial stage of great antiquity, when an ice sheet moved westward from the Continental Divide along the crest of the Sawatch Mountains, across mountain tops, such as Red Mountain, north of Aspen (*see* Aspen Quadrangle), leaving glacial drift scattered over an old surface 2000 to 3000 feet above the present drainage. Although the writer did not

examine this ancient drift, on the basis of its topographic position, it would seem to be comparable to the Cerro till of the San Juan Mountains 100 miles to the south or to the Prairie Divide stage of the northern Front Range previously described.

At Aspen, Spurr noted the prominent moraine of Roaring Fork at an elevation of 8000 feet. Examination of this moraine shows that it is comparable to the well-recognized Wisconsin II substage so prominently developed in the Upper Arkansas Valley. Moraines at a similar elevation occur on the outskirts of Aspen on both Maroon and Castle creeks. Glaciers of this substage did not coalesce at Aspen. Their outwash plains developed a prominent terrace along Roaring Fork below the town.

Spurr mentions a moraine on Roaring Fork about 5 or 6 miles above the terminal moraine at an elevation of 8000 feet. This poorly defined higher moraine is held by the writer to be the equivalent of the Wisconsin III substage, when glaciers descended from the cirque regions to elevations slightly below 9000 feet. A lateral moraine of this substage occurs across the mouth of the small valley descending from the great cirque on the north side of Independence Mountain, above the deserted mining camp of Independence. (*See* Mt. Jackson Quadrangle.)

At the mouth of Lost Man Creek is morainic debris which may belong to a later glacial substage, the Wisconsin IV. Similar morainic debris with poor topographic form lies in the main valley of Roaring Fork at an elevation of about 11,200 feet. No evidence was seen which would indicate a later glacial substage. However, some of the talus material in the cirques may be the equivalent of this last period of refrigeration.

Numerous bedrock gorges were developed in the valley of Roaring Fork after the retreat of the Wisconsin II substage and before the advance of the Wisconsin III substage, for roche moutonnée surfaces are plentiful in these narrow canyons. If the canyons had been developed before the advance of the great glaciers of the Wisconsin II substage, they should have been more severely modified and enlarged.

High bedrock spurs noted by Spurr along the upper portion of the Roaring Fork seem to represent the surface on which the oldest ice sheet moved from the mountains. Hunters Creek, which occupies a hanging valley above Roaring Fork, at Aspen, was also occupied by a large glacier which descended almost to the mouth of the creek and left a prominent terminal moraine at an elevation of approximately 8500 feet. This hanging valley does not seem to have been the product of glacial action.

Lack of time prohibited the examination of the Roaring Fork below the town of Aspen to see if the valley contained remnants of drift of the Wisconsin I substage.

Two well-developed and a possible third higher terrace lie above Aspen on the sides of Red Mountain above the outwash of the Wisconsin II glaciations. These are older than the Wisconsin glaciations, unless the lowest terrace is correlative with a Wisconsin I substage, and younger than the till on top of Red Mountain. Such a correlation is at present only speculative.

VALIDITY OF SUBDIVISION OF WISCONSIN STAGE IN SOUTHERN ROCKY MOUNTAINS

The long and continued study of the problems of European glaciation, both continental and alpine, have shown that the last stage—the Würm—which is correlated with the Wisconsin of North America was not a single ice advance and retreat. The combined attack of the geologists, paleontologists, paleobotanists, meteorologists, and archeologists has shown that there have been numerous substages of the last glacial period in Europe. The problem, however, is clearly open for greater detailed study and correlation in both Europe and North America.

In the Alps, the classic studies of Penck and Brückner (1901-1909) revealed that the last glacial stage in the Alps was divisible into at least four substages—the Würm a, the Buhl, the Gschnitz, and the Daun. Recently this subdivision has been increased until Zeuner (1935) points out that there are at least 10 substages. (*See also* Gams, 1938.) Such a subdivision is facilitated by the fact that in the piedmont region, north of the Alps, glaciers were large enough to move from the mountain valleys. Ice of later substages overrode earlier glacial and interstadial debris. Studies of stream terraces, solifluction layers, and archeological and biological remains in this region are constantly producing more information concerning the complex history of the last glacial stage.

Work of the glacial geologists and geochronologists in Northern Europe and Scandinavia indicates there is a similar complexity in the history of the last continental ice sheet. Recent studies of this intricate problem of glacial correlation are not wholly in accord with one another. A general summary and critical analysis of this problem have been presented by Zeuner (1935), Gams (1938), and Bryan and Ray (1940).

In North America, students of glacial geology are recognizing the complexity of the last glaciation. Many believe that the Iowan advance represents the earliest Wisconsin substage (Kay, 1931 a and b; Leighton, 1931) and is separated by the Peorian interstadial period from later Wisconsin advances. Leverett (1939) has recently pointed out that this belief is not universally accepted.

During the long Peorian interstadial period, the continental glacier retreated far to the north or may have completely disappeared. The

second advance of the continental ice, which formed the Harbor Hill moraine of Long Island, New York, and the Tazewell-Cary moraines of the Middle West (Bryan and Ray, 1940), was followed by a marked recession of the continental glacier or its complete but not final disappearance. In northern Wisconsin this interstadial period is marked by the Two Creeks forest bed (Wilson, 1932) in which are found the remains of a forest similar in character to that of northern Minnesota at the present time. A forest of this character indicates a marked amelioration of the glacial climate during the second Wisconsin interstadial period.

Many studies of the glacial geology and physiography of the western Cordillera hint at the possibility of numerous substages of the Wisconsin (Blackwelder, 1931; Fryxell, 1930; Horberg, 1938; Sharp, 1938), although it is generally assumed that the younger moraines mark only halts in the recession of the last ice. The complexity of river terraces, pluvial stages of the lakes of the Great Basin, and features of mountain glaciation point to marked climatic fluctuations, which in turn are reflected by advances and retreats of the mountain glaciers.

Unfortunately, the glaciers of the Southern Rocky Mountains were not large enough to extend from the mountains to the plains of the Colorado Piedmont so that there are no deposits similar to those at the foot of the Alps. However, further studies in the regions to the north, where glacial deposits of one or more of the Wisconsin substages are found in the piedmont plains, may reveal a stratigraphic sequence of glacial and interstadial deposits. Here a relationship may possibly exist between the later substages of mountain glaciation and continental glaciation. It is this northern country which may hold the key to this problem.

The reconnaissance here reported has shown that in the Southern Rocky Mountains the Wisconsin stage of glaciation is not marked by a single ice advance but is complex and can be divided into at least five substages represented by morainal deposits. The complex history, first studied in the Cache la Poudre drainage basin, is not a local phenomenon but is a regional expression of climatic fluctuations during the last stage of glaciation. Without doubt, further study of the features of the Wisconsin glaciation in the Cordilleran region will reveal that the substages here reported are not confined to the Southern Rocky Mountains but that they are the products of climatic fluctuations whose effects were of much wider extent.

In the region here described, little difficulty is encountered in correlating glacial substages from one valley to another. As expected, the moraines in general show a highly irregular but gradual increase in

elevation from north to south. The Wisconsin II moraines, for example, increase in elevation from about 7600 feet in the Cache la Poudre Canyon to 10,000 feet in the vicinity of Santa Fe, New Mexico, about 400 miles to the south (Table 3).

TABLE 3.—*Approximate elevations of Wisconsin glacial moraines in the Southern Rocky Mountains in selected valleys from southern Wyoming to northern New Mexico*

	Medicine Bow Mountains	Cache la Poudre drainage	Big Thompson drainage	Clear Creek drainage	Pikes Peak	Huerfano Valley	Red River drainage	Lake Peak
Wisc. V	—?—	11,500	11,800	—?—	—?—	11,600	12,000	—?—
IV	10,500	10,200	10,000	10,500	11,500	10,200	11,300	11,000
III	10,000	9,100–10,100	9,500	8,900–9,200	10,450–10,600	9,100	10,600 (?)	10,000
II	8,500	7,600	8,000	8,500	10,200	8,800	10,000	10,000
I	8,100	7,500 (?)	8,000 (?)	7,900	10,000	9,000	9,400	9,200 (?)

In general the moraines of the Wisconsin II, III, and IV are best preserved. Older morainal debris, which can be assigned to the Wisconsin I substage, exists in fewer places and in smaller volume. At all localities visited this older debris, where found, is more deeply weathered than that of the later substages which, because of their relative youthfulness, are almost unaltered. The Wisconsin II debris is in general only slightly iron stained and buff in the surface layers, while the Wisconsin III debris is a fresh, blue-gray boulder till. Younger tills cannot be distinguished from the Wisconsin III on the basis of weathering but must be separated primarily on the basis of topographic position.

The correlation of moraines on the basis of their topographic position is subject to the criticism that substages may have developed which have been completely obliterated by later ice advances. If such has been the case, only detailed work will reveal these substages by means other than the direct observation of morainal deposits. It was in this manner that the Wisconsin I substage was first postulated in the Cache la Poudre Valley. In the Southern Rocky Mountains the maximum extent of the Wisconsin I glaciers seems to have been almost the same as that of the Wisconsin II glaciers. In many valleys there is no trace of Wisconsin I moraines which were obliterated either by erosion or by the overriding of the Wisconsin II glaciers. However, in a few places, as in the Upper Arkansas, Clear Creek, and Huerfano valleys, remnants of the older moraines may lie in front of or in contact with the Wisconsin II moraines.

This same relationship is characteristic of the first two substages of the last glaciation in the Alps.

In the Huerfano Valley moraines of the Wisconsin I substage are not confined to the bottom of the present inner stream valleys. They lie on bedrock shoulders above the present inner valley, occupying a topographic position analogous to the Durango moraines in the San Juan Mountains. The writer suggests the possibility of an east-west axis of uplift in this region, which affected both the San Juan and the Sangre de Cristo mountains in the latitude of Sierra Blanca and the Huerfano Valley in the Wisconsin I-II interstadial period. Such a suggestion is supported by the report of deformation, some in late Pleistocene and possibly even recent time, in the San Luis Valley directly west of Sierra Blanca (Upson, 1938).

Several lines of evidence point to a long time interval between the first two Wisconsin substages. The deeper weathering of the Wisconsin I till and the excavation of the inner canyons of the San Juan Mountains and the Huerfano Valley indicate a long time interval. Direct evidence of climatic conditions during this interstadial period is lacking, except for the soil zone, about 8 inches thick, between the till of the Wisconsin I substage and the younger till of the Wisconsin II substage, reported by Wahlstrom (1939) in the valley of North Boulder Creek south of Ward, Colorado.

The major glacial and interglacial stages of the Pleistocene as well as the substages of the last Wisconsin period here reported must be explained by climatic fluctuations. Glacial advances must be the product of either increased precipitation, decreased temperature, or both. It is known that extreme coldness does not favor heavy precipitation. Because of its geographical position, it is logical to assume that the glaciation of the Southern Rocky Mountains in Wisconsin time was primarily the result of a reduction in mean annual temperature. A wetter climate was seemingly produced by reduction of the rate of evaporation.

The amount of temperature reduction needed to bring about a period of mountain glaciation in this region is unknown but it is probably only a few degrees. The effective decrease in temperature in these high mountains need be only the summer temperature, so that the snows of winter are not dissipated.

In every region there is a point above which snows are able to remain permanently. This is the climatic snow line of Hann (1903, p. 310) which ideally coincides with the lower névé limit. In the Southern Rocky Mountains the climatic snow line now lies above the mountain summits. In Pleistocene time, as now, the snow line was higher in the Southern Rocky Mountains than in the mountains of the West Coast,

where precipitation was greater. Thus, the climatic snow line in western United States may be imagined as a gently curved surface, rising gradually to the south and east, toward the center of the country, where a continental type of climate prevails.

If we assume that Hann is correct and that the climatic snow line is the lower limit of the névé, we find that the lower limit of the permanent snow fields lies above the cirque floors, for the surface of the névé is not the level of the cirque floor but lies an unknown distance above it, depending on the thickness of the glacier. If we are to reconstruct the ideal climatic snow line of the Wisconsin, the average elevation of the cirque floors must be determined, to which elevation must be added an arbitrary sum for the thickness of the ice. Exposure of the cirque and the amount and character of the highlands rising above the cirque floor must also be considered. In other words, the effects of the orographic snow line must be removed.

The orographic snow line is the lower limit of snow fields and névé which occur as isolated patches and which owe their preservation essentially to favorable mountain surroundings (Ratzel, 1886). The present fluctuating orographic snow line may be as low as 11,300 feet in the cirques on Wheeler Peak, New Mexico. Although there is no regular rise of this snow line to the south, it is generally absent in the mountains of New Mexico.

Klute (1928, p. 75) has calculated that in the Rocky Mountains the climatic snow line of the last glaciation rose from about 11,800 to 13,450 feet in the region from 41°-36° N. Lat. At present this snow line is estimated to be from 14,100 to 15,400 feet above sea level. These figures may be taken as estimates, showing the general order of magnitude. Klute believes that the snow line of the Rocky Mountains was depressed about 1960 to 2300 feet during the last glaciation—presumably he refers to the Wisconsin I or II substages. (*See* Klute, 1928, p. 75.)

Today the Southern Rocky Mountains lie an indeterminate distance below the climatic snow line. If we examine Klute's figures closely, we find that only rarely in the period of glaciation did a mountain peak in the Southern Rocky Mountains rise into the zone of the climatic snow line. Thus, we must assume that the glaciation was the result of a lowering of the orographic snow line into the zone of the mountain peaks. It follows that the lowering of the orographic snow line is dependent on the lowering of the climatic snow line.

If the assumption is made that there was more precipitation during the Wisconsin glaciation, we may expect the orographic snow line to be even further reduced in elevation. Today the orographic snow line may impress itself in the mountains at elevations about 3000 feet below the

climatic snow line proposed by Klute (1928, p. 75). If today there may occasionally be such a great difference between the climatic and orographic snow line, then, by using Klute's figures for the Wisconsin climatic snow line, we find that the orographic snow line of that time may have extended roughly to about 10,000 feet in New Mexico, or low enough to explain the development of the lowest cirques observed.

The writer is led to the belief that, in any region of mountain glaciation, it is not the climatic snow line which is effective in the formation of glaciers but the orographic, which may descend to elevations hundreds of feet below the climatic snow line.

In any reconstruction of the climate of the Wisconsin glacial substages, we must assume that the climate was cooler than the present or the interstadial periods. Once glaciers are well established in the mountains they are able to carry the colder climate down the valleys. Cold, drying winds sweep by gravity from the glacial mountain valleys on to the piedmont plains. Winds of this type were most effective on the eastern slope of the Southern Rocky Mountains, where they blew in accordance with the general prevailing winds.

On the Great Plains presumably the winds had a more northerly component than at present. The semipermanent high pressure area that existed over the continental ice sheet (Bryan and Cady, 1934, p. 244) sent a stream of cold drying winds into the central Great Plains. The westerly winds aided by those resulting from the mountain glaciation gave a westerly and perhaps a dominant component to the winds of the piedmont area. Along the mountain front, cold drying weather generally prevailed, for, even with 100 percent increase in precipitation over this area today, the climate might still be classed as semiarid to subhumid. Lowered evaporation, however, made the plains seem more moist and promoted the growth of grasses, shrubs, and mosses on the high pediments and terraces and a forest growth along the sheltered stream valleys. A somewhat analogous climate prevails today in the central portion of the Yukon River valley.

To the south and east, in Kansas, far from the ice fronts, a variable continental climate prevailed. The short, warm summers were cloudy and rainy. Winters were cold and relatively clear, with strong northwest winds. Vegetation of deciduous trees may not have survived in such a climate, for, while there was sufficient moisture, the long cold winters, similar in their effect on vegetation to prolonged droughts, and the strong winds produced a climate suitable only for coniferous forests, which doubtless flourished along the major stream valleys.

There was a cooler and wetter climate (pluvial) not only in the plains region during the glacial substages of the Wisconsin but also in the

Great Basin, where pluvial periods were coincident with the extension of the glaciers during the five substages. Numerous lakes in the undrained basins (Meinzer, 1922) attained high stands during the glacial substages and shrank during the interstadial warmer and drier periods. Recently Antevs (1935), Jones (1929), Meinzer (1922), and others have attacked the problem of the correlation of the pluvial high lake levels with the mountain glaciation, a relationship recognized by Gilbert (1890). None, however, has been entirely successful, and it is believed that the complicated series of abandoned shore lines of these former lakes will have to be re-examined in the light of a more complicated history for the last glacial stage.

The question will be raised concerning the advisability of separating the later Wisconsin into substages and abandoning the assumption that the moraines on which these substages are proposed are simply retreatal phases of the Wisconsin II glaciers. The writer believes that the substages here reported were separated from one another by interstadial periods which may have been as warm or warmer than the present. This requires that the glaciers either retreated into the cirques or disappeared completely. Such a belief is supported by the detailed study made of the terraces of the Cache la Poudre and South Platte rivers, reported by Bryan and Ray (1940).

In the Cache la Poudre Canyon, numerous terrace remnants were carefully measured and correlated. Each terrace represents the dissected surface of an outwash train developed during the maximum advance of a glacial substage. During the interstadial periods, outwash trains of the preceding substage were eroded to the canyon floor or at least below the present stream gradient. Furthermore, most of the gravel of the preceding substage was removed, so that remnants of the outwash trains are small and widely scattered. The succeeding advance of the glaciers developed an outwash plain within the re-excavated valley. As each successive substage was less extensive than the preceding, each successively younger outwash plain lies below the one immediately preceding it in origin. A terrace sequence such as this indicates successive ice advances and retreats. These ice advances were separated by relatively long periods of recession or perhaps complete disappearance of the glaciers.

CORRELATION

It has been shown that in Europe and North America there are manifold evidences of a complex series of substages of the last stage of glaciation. These suggest that climatic fluctuations of wide extent may explain these periods of glacial advance and retreat. Glaciers of the Southern

Rocky Mountains reflect these climatic fluctuations, as do those of the European Alps and the continental ice sheets of Europe and North America. An attempt to correlate the substages here reported has been made by Bryan and Ray (1940). This suggested correlation is shown in Table 4.

TABLE 4.—*Correlation of Wisconsin substages in the Southern Rocky Mountains with the Continental glaciation of North Europe and North America*

Southern Rocky Mountains	North European (continental)	North American (continental)
Sprague (Wisconsin V)	Postglacial	—?—
Long Draw (Wisconsin IV)	Fennoscanian	Cochrane (?)
?—	Scanian	—?—
Corral Creek (Wisconsin III)	Pomeranian	Late Mankato (St. Johnsbury)
Home (Wisconsin II)	Weichsel (Brandenburg)	Tazewell-Cary (Harbor Hill)
Twin Lakes (Wisconsin I)	Warthe (Fläming moraine)	Iowan (Ronkonkoma)

At present such a correlation cannot be accepted and defended dogmatically but may be considered as a marker pointing out a road which may be taken in the future after more detailed information has been gathered.

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